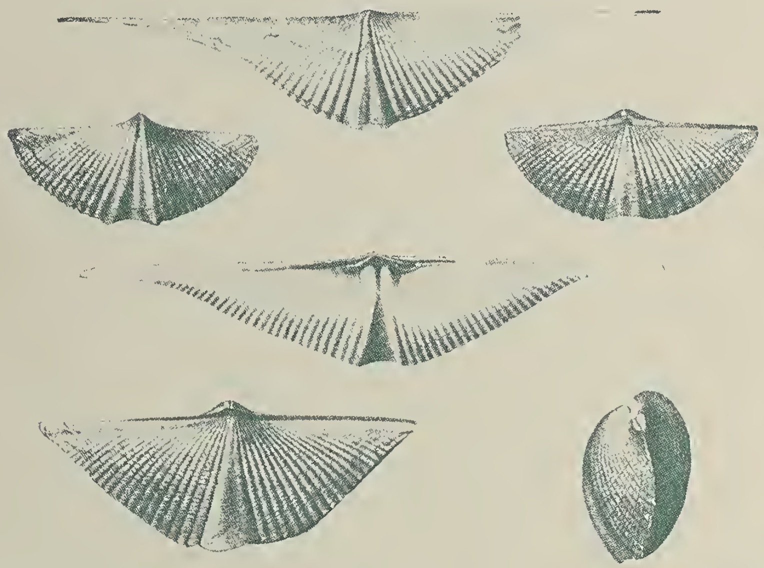
#457
Q11.N82X
NH

Q
11
N82X
NH

Dynamic Stratigraphy and Depositional Environments of the Hamilton Group (Middle Devonian) in New York State, Part I

Carlton E. Brett, *Editor*
Bulletin Number 457
New York State Museum



FEB - 5 1987

The University of the State of New York
THE STATE EDUCATION DEPARTMENT

NEW YORK
THE STATE OF LEARNING

Cultural Education Center
Albany, New York 12230

June 1986

Dynamic Stratigraphy and Depositional Environments of the Hamilton Group (Middle Devonian) in New York State, Part I

Carlton E. Brett, *Editor*
Bulletin Number 457
New York State Museum

The University of the State of New York
THE STATE EDUCATION DEPARTMENT
State Science Service

Cultural Education Center
Albany, New York 12230

June 1986

The State Education Department does not discriminate on the basis of age, color, religion, creed, disability, marital status, veteran status, national origin, race, or sex in the educational programs and activities which it operates. Inquiries concerning this policy of equal opportunity and affirmative action should be referred to the Department's Affirmative Action Officer, Education Building, Albany, NY 12234. Phone (5180 473-1265.

ISSN 0278-3355
ISBN 1-55557-003-8

THE UNIVERSITY OF THE STATE OF NEW YORK

Regents of The University *(with years when terms expire)*

1987	Martin C. Barell, *Chancellor,* B.A., I.A., LL.B., LL.D.,	Muttontown
1987	R. Carlos Carballada, *Vice Chancellor,* B.S., L.H.D., D.Sc.	Rochester
1986	Kenneth B. Clark, A.B., M.S., Ph.D., LL.D., L.H.D., D.Sc.	Hastings on Hudson
1988	Willard A. Genrich, LL.B., L.H.D., LL.D., Litt.D., D.C.S., D.C.L., Sc.D.	Buffalo
1989	Emlyn I. Griffith, A.B., J.D.	Rome
1991	Jorge L. Batista, B.A., J.D., LL.D.	Bronx
1986	Laura Bradley Chodos, B.A., M.A.	Vischer Ferry
1991	Louise P. Matteoni, B.A., M.A., Ph.D.	Bayside
1988	J. Edward Meyer, B.A., LL.B., L.H.D.	Chappaqua
1988	Floyd S. Linton, A.B., M.A., M.P.A., D.C.L., LL.D.	Miller Place
1988	Salvatore J. Sclafani, B.S., M.D.	Staten Island
1989	Mimi Lieber, B.A., M.A.	Manhattan
1992	Shirley C. Brown, B.A., M.A., Ph.D.	Albany
1990	Norma Gluck, B.A., M.S.W., LL.D.	Manhattan
1990	Thomas R. Frey, A.B., LL.B.	Rochester
1990	James W. McCabe, Sr.,A.B., M.A.	Johnson City

President of The University and Commissioner of Education
Gordon M. Ambach

Executive Deputy Commissioner of Education
Robert J. Maurer

Deputy Commissioner for Cultural Education
Carole F. Huxley

Assistant Commissioner for State Museum
Martin E. Sullivan

State Geologist and Chief Scientist, Geological Survey
Robert H. Fakundiny

Contents

The Middle Devonian Hamilton Group of New York: An Overview
Carlton E. Brett 1

Redefinition, Stratigraphy and Depositional Environments of the Mottville Member (Hamilton Group) in Central and Eastern New York
Thomas X. Grasso 5

Faunal and Lithologic Cyclicity in the Centerfield Member (Middle Devonian, Hamilton Group) of Western New York: A Reinterpretation of Depositional History
Michael Savarese, Lee M. Gray and Carlton E. Brett 32

Depositional Environments and Sequences, *"Pleurodictyum Zone"*, Ludlowville Formation of Western New York
Keith B. Miller 57

Deposition and Early Diagenesis of A Middle Devonian Marine Shale: Ludlowville Formation, Western New York
Glenn T. Wygant 78

Petrology, Taphonomy and Sedimentary Environments of Pyritic Fossil Beds from the Hamilton Group (Middle Devonian) of Western New York
Vincent B. Dick and Carlton E. Brett 102

Storm-Generated Sedimentary Units: Tempestite Proximality and Event Stratification in the Middle Devonian Hamilton Group of New York
Carlton E. Brett, Stephen E. Speyer and Gordon G. Baird 129

Acknowledgments

Acknowledgment is made to the National Science Foundation (NSF grant EAR-8313103) and to the Donors of the Petroleum Research Fund, administered by the American Chemical Society, for partial support of this publication.

THE MIDDLE DEVONIAN HAMILTON GROUP OF NEW YORK: AN OVERVIEW

Carlton E. Brett
Department of Geological Sciences
University of Rochester

The Middle Devonian Hamilton Group of New York State comprises one of the most complete and fossiliferous sequences of middle Paleozoic strata in the world. Hamilton strata represent the first major phase of terrigenous sedimentation from the Acadian orogenic belt — a prominent change from the clean carbonates of the underlying Onondaga Limestone and a precursor to the Late Devonian Catskill Deltaic sequence. This unit forms an eastwardly thickening and coarsening wedge of marine shales, siltstones, and sandstones, punctuated by thin, but persistent, carbonate beds (see Cooper, 1957; Rickard, 1975, 1981).

Ranging from just under 85 m (280 feet) at Lake Erie to over 900 m (3000 feet) on the Catskill Front, the Hamilton strata span latest Eifelian to middle Givetian age, recording approximately 5 to 7 million years of geologic history in the northern Appalachian Basin (Rickard, 1975). Direct Rb-Sr dating of lower Hamilton black shales has produced dates ranging between 377 ± 11 and 384 ± 9 million years (Bofinger and Compston, 1967).

The extraordinarily well preserved and diverse invertebrate fossils of the Hamilton Group, including corals, brachiopods, trilobites, and echinoderms, were documented in detail through the classic studies of James Hall in the mid to late 1800's (see Fisher, 1981, for historical account). The term Hamilton was introduced to the literature by Vanuxem (1840) based on classic exposures in the Chenango River Valley, near Hamilton, Madison County, New York. Subsequently, Hamilton strata were subdivided, largely on the basis of thin limestone marker beds, into four formations (ascending order): the Marcellus, Skaneateles, Ludlowville, and Moscow.

In the early 1900's monumental stratigraphic and facies studies by Cleland (1903), Chadwick (1933), Smith (1916, 1935), and, above all, G. Arthur Cooper (1930, 1933, 1934) provided a detailed framework of Hamilton lithostratigraphy and general facies relationships. There then followed an interval of thirty years, during which time various paleontologists updated the systematics of particular fossil groups and numerous workers refined the stratigraphic relationships (Goldring, 1935, 1943; Rickard, 1952; Oliver, 1951; Rickard and Zenger, 1964). However, little comprehensive paleoenvironmental study was undertaken.

In the past 20 years, interest in the Hamilton has been rekindled by comparative studies of Upper Devonian facies (Sutton, et al., 1970; Bowen et al., 1974; Thayer, 1974) and by the research of Friedman and Johnson (1968), McCave (1969a, 1969b, 1973), Heckel (1973), Grasso (1970, 1973), Baird (1978, 1979), and others, on facies models and paleoecology. The present symposium volume is one outcome of this resurgence of research interest in Hamilton paleoenvironments.

Given its abundant and well preserved fossils, good surface exposures and detailed stratigraphy, the Hamilton provides an excellent framework for the development and testing of models relating to sedimentology, paleoecology, biogeography and evolutionary tempo and mode. Surprisingly, this sequence has been largely overlooked with regard to such work, with a few notable exceptions. For instance, the evolutionary model of punctuated equilibrium was founded, in part, on Niles Eldredge's observations of the Hamilton trilobite *Phacops rana* (Eldredge and Gould, 1972). However, on the whole, paleobiologic and paleoenvironmental studies of the Hamilton Group have lagged behind those of other classic fossil-bearing strata.

Devonian sedimentary rocks of New York State have proven to be a major source of data on depositional environments and fossil communities in ancient epeiric seas. However, to date, most of the detailed paleoenvironmental studies have concentrated either on Lower and Middle Devonian carbonate shelf sequences (see Rickard, 1962; Laporte, 1969, for Helderberg; Koch, 1981; Feldman, 1980, for Onondaga) or on the Upper Devonian Catskill deltaic complex (Sutton et al., 1970; Bowen et al., 1974; Thayer, 1974; McGhee and Sutton, 1981). Reefs and other carbonate facies of the Middle Devonian Onondaga and Tully Limestones, which, respectively, underlie and overlie the Hamilton Group, have also received recent detailed study particularly with regard to their economic potential (Mesolella, 1978; Feldman, 1980; Lindeman, 1979; Lindeman and Feldman, 1981; Koch, 1981; Kissling and Moshier, 1981; Heckel, 1973). In contrast, intervening Hamilton strata have been the focus of relatively little study (but see Grasso, 1970, 1973; Grasso and Wolf, 1977; Wolff and Buttner, 1980; McCave 1969a, 1969b, 1973). The trite phrase "warm, shallow, muddy seas" nearly sums up what most paleontologists have had to say about Hamilton depositional environments. Why is this so?

Several factors may have impeded Hamilton paleoecologic/paleoenvironmental studies. Firstly, there is the problem of familiarity. Hamilton fossils were among the first described in North America and the general framework of lithostratigraphy has long been known. Because Middle Devonian rocks are exposed in an easily accessible area there is a strong tendency, particularly among new workers and students, to assume that the rocks have been overstudied. However, as pointed out, this is not true. The Hamilton offers an abundance of challenging problems and has presented us with many surprises.

Secondly, there is a real problem concerning biostratigraphy. The Givetian Stage, on the whole, is not easily zoned; traditionally-used index fossils (i.e., goniatitic ammonites and conodonts) display very little evolutionary change during this time interval. For example, the upper half of the Hamilton Group (Ludlowville and Moscow Formations) lies within part of a single goniatite zone (characterized by *Tornoceras uniangulare;* cf.House, 1965, 1981; Rickard, 1975) and in the lower and middle subzones of the *Polygnathus varcus* conodont zone (Ziegler et al., 1976; Klapper, 1981). Clearly, stratigraphers must find alternative means of establishing more detailed temporal relationships. We are presently investigating two such techniques: location of cycle centers (facies reversal planes) and mapping the distribution of single event beds (e.g. widespread storm layers). Ironically, certain traditional assumptions made by past Hamilton stratigraphers may turn out to be right for the wrong reasons. For example, the "Centerfield Member" (*sensu lato*) may be a timeparallel unit, but not because it reflects widespread, synchronous limestone deposition over the entire basin, as has often been assumed. In fact, as previously defined, the term Centerfield Member does not denote a single lithological unit, but rather stands for a broad, cyclic depositional package which includes coral-rich limestones, sparsely fossiliferous bioturbated mudstones, and even cross bedded sandstones (Gray, 1983). In outcrop the "Centerfield" forms waterfalls, thus being easily recognizable from background shales. Herein lies the critical link to time; the resistant middle beds of the Centerfield lie at the center of a subsymmetrical sedimentary sequence. What actually has been mapped is the reversal plane of a facies cycle; in this sense the falls capping unit in any given outcrop marks an approximate chronostratigraphic unit. The Centerfield limestone *sensu stricto* is only one component facies of the Centerfield cycle and it may well be slightly diachronous in its three-dimensional extent.

The third difficulty may be referred to as the "layer cake stratigraphy problem." In western New York stratigraphic sequences in the Hamilton include numerous, widely persistent thin horizons that display very little lateral facies change. Although this layer cake aspect of western Hamilton stratigraphy facilitated regional correlations, it has sometimes fostered a non-Waltherian view of vertically stacked Hamilton facies, suggesting simultaneous, basinwide environmental changes. We have all learned to mistrust layer cake patterns in stratigraphy, generally; consequently, certain workers have tended to suspect that the pattern of continuous beds in the Hamilton as more an artifact of stratigraphers' imagination than reality. I wish to emphasize that this is most assuredly not the case. In fact, the layer cake pattern is an important phenomenon which we must account for in any explanation of the Hamilton Group or of other, similar, deposits. To paraphrase Stephen Gould, layer cake stratigraphy 'is data' and should not be ignored or argued away.

I believe there also has been a problem of misinterpreting the temporal scale of many Hamilton beds. Detailed studies of Hamilton units indicate that many exhibit features indicative of rapid single event deposition while others yield evidence for long and complex sedimentary histories. Such disparities in scale must be evaluated to fully decipher depositional sequences in the Hamilton.

Finally, I visualize an odd problem, which I refer to as the "allure of the Catskill delta." The juxtaposition of the Late Devonian Catskill delta on top of the Hamilton has, I think, had an undue influence on the interpretation of Hamilton beds. First, sedimentologists and paleontologists have tended to focus on the better developed deltaic facies of the Upper Devonian because, indeed, the Upper Devonian does fit an easily conceptualized deltaic model better than the Hamilton does. Conversely, Hamilton workers have tended to extrapolate, wholesale, the delta model to the Hamilton, even though the fit is often strained, at best. The carbonate shelf/ramp/basin model of the Onondaga Limestone recently proposed by Kissling and Moshier (1981), Koch (1981) and others may actually provide a more reasonable explanation for Hamilton facies distribution in western New York. The best facies model for the Hamilton Group overall will probably be a hybrid of the Onondaga carbonate shelf and the Catskill deltaic complex models (see Rickard, 1975).

Not surprisingly, delta enthusiasts have tended to concentrate their efforts on eastern siliciclastic facies; the relatively few sedimentological studies of the Hamilton Group (McCave, 1967, 1969a, 1969b, 1973; Mazullo, 1973; Wolf and Buttner, 1980) have been largely confined to paralic depositional environments of the Catskill region. In contrast, paleontologists have tended to focus on the paleoecology of the fossil-rich western New York strata (Beerbower, et al., 1969; Grasso, 1970, 1973). Little attempt has been made to relate or integrate these disparate types of interpretation. The key intermediate area, roughly between Seneca Lake and the Unadilla Valley, must be studied in detail to resolve these two end points; this region is the focus of much of our current research.

The present volume largely represents the results of research first presented at a Paleontological Society Symposium, held March 24, 1983, at the Northeast Section meeting of the Geological Society of America in Kiamesha Lake, New York. The papers reflect a range of research including microstratigraphic-

and paleoenvironmental studies of particular units (papers by Grasso, Savarese, et al., and Miller), as well as more broad-based sedimentologic and taphonomic surveys (Wygant, Dick and Brett, and Brett, et al.). However, all of the papers involve process-oriented analysis of sedimentary sequences and hence are united in their focus on "dynamic stratigraphy" (sensu Matthews, 1984) and depositional environments.

I hope that this volume will demonstrate that there is now a good deal of research activity on various aspects of the Hamilton and that many of the long-ignored problems are now being tackled. I also hope this symposium will stimulate further interest in the study of various aspects of Hamilton stratigraphy, sedimentary environments, and paleobiology. It is fitting that this volume be dedicated to G. Arthur Cooper, whose pioneering work on Hamilton facies fostered so much of what is to come.

References Cited

Baird, G.C. 1978. Pebbly phosphorites in shale: a key to recognition of a widespread submarine discontinuity in the Middle Devonian of New York. J. Sed. Petrol. 48:545-555.

_________ 1979. Sedimentary relationships of Portland Point and associated Middle Devonian rocks in central and western New York. New York State Mus. Bull. 433, 24 p.

Beerbower J.R., Buehler, E.J., Bray, R.G., and Jordon, F.W. 1969. A paleomicroecologic study of Mid Devonian marine assemblages (Hamilton Group) in western New York. North American Paleontological Convention, Field Trip Guidebook 5:48-78.

Bofinger, J.M. and Compston, W. 1967. A reassessment of the age of the Hamilton Group, New York and Pennsylvania, and the role of inherited radiogenic Sr^{87}. Geochim. et Cosmochim. Acta 31:2353-2359.

Bowen, Z.P., Rhoads, D.C., and McAlester, A.L. 1974. Marine benthic communities of the Upper Devonian of New York. Lethaia 7:93-120.

Chadwick, G.H. 1933. Great Catskill delta: and revision of the Late Devonic succession. Pan.-Amer. Geol. 60:91-360.

Cleland, H.F.1903. A study of the fauna of the Hamilton Formation of the Cayuga Lake section in central New York. U.S. Geol. Surv. Bull. 206, 112 p.

Cooper, G.A. 1930. Stratigraphy of the Hamilton Group of New York. Amer. J. Sci., 5th Ser., 19:116-134, 214-236.

_________. 1933, 1934. Stratigraphy of the Hamilton Group of eastern New York, Pts. 1,2. Amer. J. Sci. 26:537-551; 27:1-12.

_________ 1957. Paleoecology of the Middle Devonian of eastern and central United States. *In* Ladd, H.S., ed., Treatise on Marine Ecology and Paleoecology, Pt. 2, Paleoecology. Geol. Soc. Amer. Mem. 67:249-278.

Eldredge, N. and Gould, S.J. 1972. Punctuated equilibrium: an alternative to phyletic gradualism. *In* Schopf, T.J.M., ed., Models in Paleobiology. Freeman, Cooper & Co., San Francisco. p.82-115.

Feldman, H.R. 1980. Level-bottom brachiopod communities in the Middle Devonian of New York. Lethaia 13:27-46.

Fisher, D.W. 1981. History of Devonian geology in New York. *In* Oliver, W.A., Jr. and Klapper, G., eds., Devonian Biostratigraphy of New York. Intl. Union Geol. Sci., Subcomm. on Devonian Strat., Washington, D.C., p.23-52.

Friedman, G.M. and Johnson, K.G. 1968. The Devonian Catskill deltaic complex of New York, type example of a tectonic delta complex. *In* Shirley, M.L. and Ragdale, J.A., eds., Deltas in their Geologic Framework, Houston Geol. Soc., Houston, Texas, p.189-211.

Goldring, W. 1935. Geology of the Berne Quadrangle. New York State Mus. Bull. 303, 374 p.

_________ 1943. Geology of the Coxsackie Quadrangle. New York State Mus. Bull. 332, 208 p.

Grasso, T.X. 1970. Paleontology, stratigraphy, and paleoecology of the Ludlowville and Moscow Formations (upper Hamilton Group) in central New York. *In* New York State Geol. Assn. Guidebook, 42nd Ann. Mtg., SUNY College at Cortland, p. D11-22.

_________ 1973. A comparison of environments. The Middle Devonian Hamilton Group in the Genesee Valley. *In* New York State Geol. Assn. Guidebook, 45th Ann. Mtg., SUNY College at Brockport, p. B1-B23.

_________ and Wolff, M.P. 1977. Paleoenvironments of the Marcellus and lower Skaneateles Formations of the Otsego County region (Middle Devonian). *In* New York State Geol. Assn. Guidebook, 49th Ann. Mtg., SUNY College at Oneonta, p. 1-39.

Gray, L.M. 1983. Biofacies and depositional environments of the Centerfield "Member" in the northern Appalachian Basin. Geol. Soc. Amer. Abstr. with Programs 15(3):170.

Heckel, P.H. 1973. Nature, origin, and significance of the Tully Limestone. Geol. Soc. Amer. Spec. Pap. 138, 244 p.

House, M.R. 1965. A study in the Tornoceratidae: the succession of *Tornoceras* and related genera in the North American Devonian. Phil. Trans. Roy. Soc. 250B:79-130.

_________ 1981. Lower and Middle Devonian goniatite biostratigraphy. In Oliver, W.A., Jr. and Klapper, G., eds., Devonian Biostratigraphy of New York. Intl. Union Geol. Sci., Subcomm. on Devonian Strat., Washington, D.C., p. 33-37.

Kissling, D.L. and Moshier, S.U. 1981. The subsurface Onondaga Limestone: stratigraphy, facies and paleogeography. In Enos, P., ed., New York State Geol. Assn. Guidebook, 53rd Ann. Mtg., SUNY at Binghamton, p. 279-280.

Klapper, G. 1981. Review of New York conodont biostratigraphy. In Oliver, W.A., Jr. and Klapper, G., eds., Devonian Biostratigraphy of New York. Intl. Union of Geol. Sci., Subcomm. on Devonian Strat., Washington, D.C., p. 57-66.

Koch, W.F. 1981. Brachiopod community paleoecology, paleobiogeography, and depositional topography of the Devonian Onondaga Limestone and correlative strata in eastern North America. Lethaia 14(2):83-103.

Laporte, L.F. 1969. Recognition of a transgressive carbonate sequence within an epeiric sea: Helderberg Group (Lower Devonian) of New York State. *In* Friedman, G.M., ed., Soc. Econ. Paleontol. and Mineral. Spec. Pub. 14:98-119.

Lindemann, R.H. 1979. Stratigraphy and depositional history of the Onondaga Limestone in eastern New York. New York State Geol. Assn. Guidebook, 51st Ann. Mtg., Rensselaer Polytech. Inst., p. 351-385.

__________ and Feldman, H.R. 1981. Paleocommunities of the Onondaga Limestone (Middle Devonian) in central New York State. New York State Geol. Assn. Guidebook, 53rd Ann. Mtg., SUNY at Binghamton, p. 79-93.

Matthews, R.K. 1984. Dynamic Stratigraphy. Prentice-Hall, New Jersey, 2nd ed., 489 p.

Mazullo, S.J. 1973. Deltaic depositional environments in the Hamilton Group (Middle Devonian), southeastern New York State. J. Sediment. Petrol. 43(4):1061-1071.

McCave, I.N. 1967. A stratigraphical and sedimentological analysis of a portion of the Hamilton Group (Middle Devonian) of New York State. Doctoral dissertation, Brown Univ., 381 p.

__________ 1969a. Correlation of marine and non-marine strata with an example from the Devonian of New York. Amer. Assn. Petrol. Geol. Bull. 53:155-162.

__________ 1969b. Correlation using a sedimentological model of part of the Hamilton Group (Middle Devonian) of New York. Amer. J. Sci. 267:567-591.

__________ 1973. The sedimentology of a transgression: Portland Point and Cooksburg Members (Middle Devonian) New York State. J. Sediment. Petrol. 43(2):484-504.

McGhee, G.R. and Sutton, R.G.1981. Late Devonian marine ecology and zoogeography of the central Appalachians and New York. Lethaia 14:27-43.

Mesolella, K.J. 1978. Paleogeography of some Silurian and Devonian reef trends in the central Appalachian Basin. Amer. Assn. Petrol. Geol. Bull. 62:1607-1644.

Oliver, W.A., Jr. 1951. Middle Devonian coral beds of central New York. Amer. J. Sci. 249:705-728.

Rickard, L.V. 1952. The Middle Devonian Cherry Valley Limestone of eastern New York. Amer. J. Sci. 250:511-522.

__________ 1962. Late Cayugan (Upper Silurian) and Helderbergian (Lower Devonian) stratigraphy in New York. New York State Mus. Bull. 386, 157 p.

__________ 1975. Correlation of the Silurian and Devonian rocks in New York State. New York State Mus. Map and Chart Ser. 24, 16 p.

__________ 1981. The Devonian System of New York State p.5-21. *In* Oliver, W.A., Jr., and Klapper, G., eds., Devonian Biostratigraphy of New York. Intl. Union Geol. Sci. Subcomm. on Devonian Strat., Washington, D.C., p. 5-21.

__________ and Zenger, D.H. 1964. Stratigraphy and paleontology of the Richfield Springs and Cooperstown Quadrangles, New York. New York State Mus. Bull. 396, 101 p.

Smith, B. 1916. The structural relations of some Devonian shales in central New York. Acad. Nat. Sci. Phila. Proc. 67:561-569.

__________ 1935. Geology and mineral resources of the Skaneateles Quadrangle. New York State Mus. Bull. 300, 120 p.

Sutton, R.G., Bowen, Z.P., and McAlester, A.L. 1970. Marine shelf environments of the Upper Devonian Sonyea Group of New York. Geol. Soc. Amer. Bull. 81:2975-2992.

Thayer, C.W. 1974. Marine paleoecology in the Upper Devonian of New York. Lethaia 7:121-155.

Vanuxem, L. 1840. Fourth annual report of the Geological Survey of the Third District. New York Geol. Surv. Ann. Report 4.

Wolff, M.P. and Buttner, P.J.R. 1980. Marine and fluvial delta platform environments of the transgressive clastic correlatives of the Middle Devonian (Erian) Mottville Limestone Member of the Skaneateles Formation in eastern New York State. New York State Geol. Assn. Guidebook, 52nd Ann. Mtg., Rutgers, p. 243-271.

Ziegler, W., Klapper, G.,and Johnson, J.G. 1976. Redefinition and subdivision of the *varcus* Zone (Conodonta, Middle-?Upper Devonian) in Europe and North America. Geol. et Palaeontol. 10:109-140.

Redefinition, Stratigraphy and Depositional Environments of the Mottville Member (Hamilton Group) in Central and Eastern New York

Thomas X. Grasso
Geoscience Department
Monroe Community College

ABSTRACT

The Mottville Member (base of the Skaneateles Formation) is redefined to encompass, at the type locality approximately 20 feet (6m) of hard, calcareous mudstone, shale, and a basal band (1 foot thick [0.3m]) of dark impure limestone, not 50 feet (15m) as originally defined by Smith. The unit thickens eastward from Cayuga Lake to 50 feet (15m) in the Cazenovia region where two limestone beds are separated by a wedge of nearly barren shale attaining a maximum thickness of 40 feet (12m). The upper limestone unit is bracketed above and below by relatively thin, geographically persistent, fossiliferous shales. The lower limestone bed and the barren shales pass into the upper Peckport Member in the Chenango Valley while the upper limestone and its associated fossiliferous, shales grade into 10 to 20 feet (3 to 6m) of sandy coquinites that continue east of the Unadilla Valley. Four fossil communities are recognized in the Mottville: *Tropidoleptus* Community below the upper limestone bed, *Heliophyllum-Favosites* Community within the upper limestone, *Stereolasma* Community above the upper limestone, *"Camarotoechia"-Allanella* Community in the sandy coquinites. The west to east distribution of communities and lithologies suggests an offshore to onshore depositional setting. The Mottville and other Hamilton carbonates that serve as formational boundaries have been interpreted as representing brief episodes of transgressive maxima. The present study suggests that the vertical succession of litho- and biofacies form an upward shallowing sequence. The entire Mottville lies between deeper water deposits of the Cardiff and its equivalents below and the Delphi Station and its equivalents above. Therefore, the Mottville time-equivalent in the Catskill red bed facies should be found at or near the top of a regressive tongue, not a transgressive one.

INTRODUCTION

The Devonian section in New York State begins with carbonates (Helderberg Group and Onondaga Limestone) deposited during Early and Middle Devonian time, which are overlain by a thick, upward coarsening, regressive, clastic sequence that represents a westward migrating deltaic complex (Catskill Delta) constructed during late Middle and Late Devonian time.

The Middle Devonian Hamilton Group rests upon the Onondaga Limestone and records the initial clastic influx of the Catskill wedge. The Hamilton is 265 feet thick (81m) on Lake Erie and thickens eastward to 640 feet (195m) at Canandaigua Lake, 1100 feet (335m) in the Tully Valley, 1750 feet (533m) in the Unadilla Valley, 2000 feet (610m) in the Susquehanna Valley and 2200 feet (671m) near Richmondville (Rickard, 1975, Plate 4). The highly fossiliferous Hamilton beds are well exposed, structurally uncomplicated and lend themselves to detailed stratigraphic, paleontologic and paleoecologic studies. Although Lardner Vanuxem (1842) and James Hall (1843) in their monumental reports on the third and fourth geological districts, respectively, laid the foundation for all later work, the general stratigraphic relationships of the Hamilton Group as presently understood were first described in Cooper's classic papers (1930, 1933). Within the Hamilton Group there are three geographically extensive carbonate horizons that serve as time-planes or keybeds. These relatively thin units do not change facies as rapidly as the units above and below and they penetrate the clastic wedge, subdividing it into four units, in ascending order: the Marcellus, Skaneateles, Ludlowville and Moscow Formations. The lowest of these carbonate units, the Stafford Member (west of Cayuga Lake) and Mottville Member (east of Cayuga Lake) marks the base of the Skaneateles Formation and separates it from the underlying Marcellus Formation (Figure 3). The other Hamilton carbonate horizons that serve as formational boundaries are the Centerfield Member, between the Skaneateles and Ludlowville Formations, and the Tichenor (western New York) — Portland Point Members (central and eastern New York), between the Ludlowville and Moscow Formations.

Because the carbonate horizons occur in a predominately clastic sequence, the assumption has been made that they represent brief episodes of widespread transgressions. The stud-

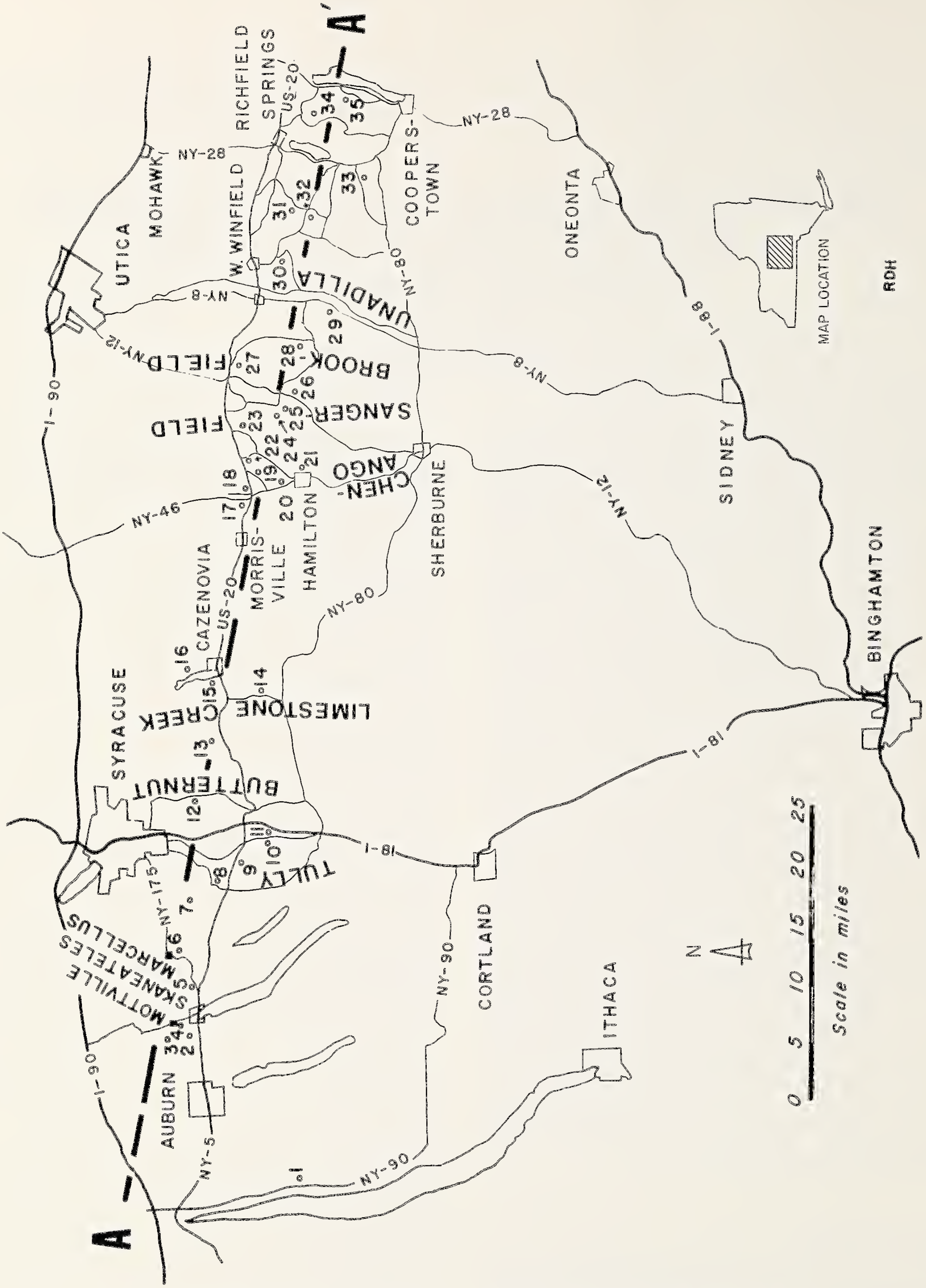


Figure 1. Distributions of outcrops studied. Numbered outcrops are listed and described in Appendix I.

ies of Brett and Baird (1982), Baird and Brett (1983), Brett and others (1983), Gray (1983), Selleck (1983), and others indicate that these, as well as other Hamilton units, are parts of shallowing upward or regressive sequences. As will be shown here, this latter model applies to the Mottville Member.

In this report the Mottville is redefined on lithologic and faunal criteria to strata that are mappable from Cayuga Lake to at least the Unadilla Valley and perhaps the west side of Otsego Lake, a distance of 95 miles. It will be shown that the lower and upper contacts of the newly defined Mottville are readily apparent in the field and allow more precise correlations to be made within the Skaneateles Formation across central New York. The fossil communities and trophic structures of the Mottville are also described. Finally, the paleoenvironmental settings which prevailed before, during and immediately following Mottville deposition are discussed.

METHODS AND PROCEDURES

Thirty-five localities of the Mottville were measured and sampled between the east side of Cayuga Lake and the vicinity of Cooperstown (Figure 1). Appendix 1 lists the localities and exposed units. At each locality lithologic and faunal collections were made from the Mottville Member and from strata immediately above and below. Samples from certain levels were thin sectioned and examined to verify field identifications. The relative abundance of each fossil taxon was established in the field by estimating the number of specimens observed in one hour from each richly fossiliferous horizon. The abundance of each taxon was then assigned a code number as follows: (1) 1 to 3 specimens, (2) 4 to 12 specimens, (3) 13 to 50 specimens, (4) 50 to 200 specimens, (5) 200 to 800 specimens, (6) over 800 specimens. These data are shown in Appendix II.

REDEFINITION OF MOTTVILLE MEMBER

The Mottville Member, as originally defined by Smith (1916, p. 562) in the Skaneateles 15 minute quadrangle, included about 20-25 feet (7 m) of shale capped by 1 foot (0.3 m) of limestone and a few feet of moderately hard shale. Later, Smith (1935) added to this another thin limestone bed a few feet above the lower, overlain by another 15 to 20 feet (8 m) of hard calcareous shale, thereby expanding the Mottville to include 40 to 50 feet (12 to 15 m) of strata. Smith designated as the type section, the cut along the railroad spur and on both sides of Skaneateles Creek, south of the village of Mottville. Only the upper limestone bed continues to the west edge of the Skaneateles Quadrangle, and by inference to Cayuga Lake, while the lower limestone is exposed only along the east margin of the quadrangle. Figure 2a, extracted from Smith (1935, Fig. 16, p.38), illustrates his concept of the Mottville and its stratigraphic relationships. Smith's definition was based entirely on faunal evidence, as he believed the Mottville to be a highly fossiliferous zone rapidly transitional from the Cardiff below, "...beginning with *Leiorhynchus* and its associates but later becoming a *Phacops-Homalonotus-Spirifer* assemblage" (Smith, 1935, p.12).

Cooper (1930, p. 217) states:

> "The Mottville at the type locality is composed of soft shale and harder calcareous shale, in the midst of which is a layer of hard limestone about one foot thick, which not uncommonly contains much crinoid debris and corals."

Cooper correlated the hard calcareous shales of the Limestone Creek Valley-Cazenovia region with Smith's upper hard shales of the type locality in the Skaneateles Quadrangle.

The Mottville as defined above is too impractical and imprecise for accurate field mapping because the lower and upper contacts could range through many feet of strata from one locality to the next. Furthermore, as a key bed the unit needs to be delimited as precisely as possible. More specifically, it should be recognized and mapped primarily on the basis of its lithology, namely the thin limestone beds, and subordinately the fauna. This approach was tested by mapping the Mottville across the Skaneateles Quadrangle, placing its lower contact at the base of the lower carbonate bed. The results, superimposed on Smith's illustration (Fig. 2a), are shown in Figure 2b and discussed in greater detail in the next section.

The general stratigraphic relationships of the Mottville Member are shown in Figure 3. The underlying Marcellus Formation is approximately 100 feet (30 m) thick at Cayuga Lake, thinning to 80 feet (24 m) near Mottville, then gradually thickening eastward to nearly 800 feet (244 m) in the Susquehanna Valley. The Marcellus Formation thicknesses are from Fernow (1961) for the Cayuga Lake meridian and Cooper (1930, 1933) and Rickard (1975) for those meridians east of the Tully Valley; the remainder are from the author's field measurements. The Mottville is 8.5 feet (2.6 m) thick at Cayuga Lake, 20 feet (6 m) thick at the type locality, thins to about 5 feet (1.5 m) near Marcellus and thickens again eastward to 50 feet (15 m) in the Cazenovia 15 minute quadrangle. East of the Chenango Valley to Richfield Springs only the upper 10 to 20 feet (3 to 6 m) of the unit are present, as the lower and middle parts of the Mottville grade eastward into the Pecksport Member.

The Mottville overlies the Oatka Creek Member (black shale) at Cayuga Lake, the Cardiff Member (dark gray, clay shale) from just west of Mottville eastward to the vicinity of Morrisville, and the Pecksport Member (dark gray and silty shales) from there to the Richfield Springs — Cooperstown region. The top of the Solsville Member rises eastward relative to the Mottville as the Pecksport thins from approximately 165 feet (50 m) in the Chenango Valley to 90 feet (27 m) in the Unadilla Valley and 60 feet (18 m) west of Richfield Springs. The Mottville may indeed merge with the Solsville east of Otsego Lake.

The Levanna Member (black shale) of the Skaneateles Formation overlies the Mottville at Cayuga Lake but is replaced eastward by dark gray, clay shales and then dark silty shales of the Delphi Station Member.

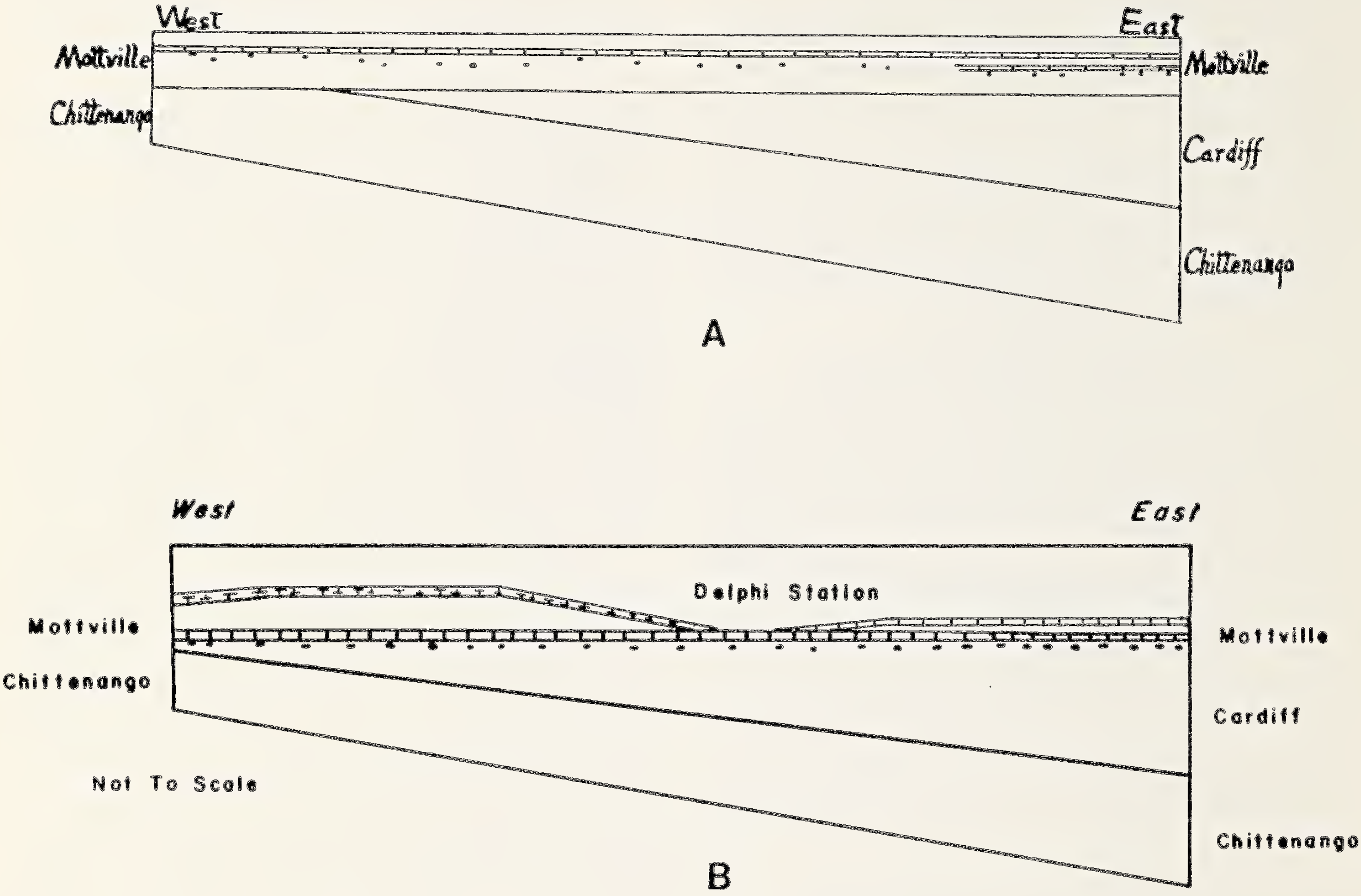


Figure 2a. Diagrammatic section showing relations of Smith's (1935) Chittenango, Cardiff and Mottville Members on the Skaneateles Quadrangle. Member contacts are transitional. Not drawn to scale.

2b. Redefined Mottville superimposed on Smith's (1935), Figure 16 for the Skaneateles Quadrangle. Not drawn to scale.

STRATIGRAPHY OF MOTTVILLE

As described above, the Mottville can be more precisely defined on the basis of its lithology and contained fauna. A detailed cross section of the redefined Mottville, with the sections projected to the vertical plane A-A' of Figure 1 at 90°, is shown in Figure 4. Lithologically there is no distinct break above the Cardiff or Pecksport Members until the first limestone bed of the Mottville is encountered, a unit hereafter referred to as limestone "a," which is Smith's lower limestone bed. This unit can be readily traced from Locality 2 (Fig. 1) west of Mottville to Locality 18 near Pine Woods (Chenango Valley), a distance of approximately 50 miles. West of Locality 2 to near Cayuga Lake there is a gap of nearly 15 miles where no exposures of the Mottville could be found. The author believes the lower 8 inches (20 cm) of the Mottville at Cayuga Lake to be limestone "a". The remaining 8 or so feet (2 m) of the Mottville correlates with the uppermost part of the Mottville to the east.

A thin bed, 2 to 3 feet thick (0.9 m), containing enormous numbers of the brachiopods *Ambocoelia, Leiorhynchus,* chonetids, and the small, thin shelled, bivalve *Pterochaenia fragilis* is present approximately 13 feet (4 m) below limestone "a" at Locality 2 and 15 feet (4.5 m) below at Locality 5 east of Mottville (Fig. 4). Approximately 13 feet (4 m) below limestone "a" at Great Gully (locality 1) an *Ambocoelia* bed containing *Leiorhynchus,* chonetids and *Pterochaenia* is exposed. The author believes this to be the same *Ambocoelia* bed mapped to the east. If this correlation is correct the base of the Mottville at Great Gully is limestone "a." At Great Gully, limestone "a" (20 cm) contains a fauna of small brachiopods (*Ambocoelia* and chonetids), many possessing a bluish sheen, trilobites (*Dipleura, Phacops*) and nautiloids similar to limestone "a" at Locality 2.

Defining the lower contact of the Mottville as the base of limestone "a" also is supported by the fauna. The Mottville, throughout most of the study area, does not become a highly fossiliferous zone, as Smith stated, until the base of limestone "a" where the number of species as well as individuals increases significantly from the Cardiff-upper Pecksport below. These Marcellus units generally contain a low diversity assemblage of *Ambocoelia,* chonetids, and *Leiorhynchus,* although locally, as in the Tully Valley, the Cardiff gradually increases in diversity upwards, embracing a greater number of species near the top. The upper contact of the Mottville with the Delphi Station Member is marked by the sudden disappearance of most taxa from within the unit, save for *Ambocoelia,* chonetids, and gastropods such as *Bembexia*. Although this contact is not as obvious lithologically as the lower one, it nevertheless is easily recognizable in the field.

On the east side of Cayuga Lake (Locality 1) the Mottville consists of an 8 inch (20 cm) fine grained, nodular, dark gray, fossil-rich calcareous band, equivalent to limestone "a", overlain by about 8 feet (2 m) of bioturbated, dark gray, argillaceous limestone. *Zoophycos* traces are the only organic remains present in this upper interval. At Mottville (Locality 4) limestone "a" is about 1 foot (0.3 m) of dark gray, thin bedded limestone, with stringers of crinoidal material, containing a moderately diverse assemblage of small brachiopods, trilobites (*Dipleura, Phacops*), nuculid bivalves, and gastropods. This unit is overlain by about 20 feet (6 m) of hard calcareous, sparsely fossiliferous mudstone, probably equivalent to the argillaceous limestone overlying limestone "a," at Great Gully. Mottville faunal elements are detailed in the next section of this paper and in Appendix 2.

Four miles east of Mottville at Falls Ravine (Locality 5) the Mottville Member is only 3 feet thick (1 m) and consists of 2 feet (0.6 m) of gray calcareous shale, containing a moderately diverse assemblage rich in the small rugose coral *Stereolasma* and the trilobite *Phacops rana,* overlying limestone "a". This locality and another small gully to the northeast are significant for the first appearance of *Stereolasma*-rich shales east of the type locality. The top of these shales marks the upper contact of the Mottville because the dark shales of the Delphi Station Member immediately above contain a very low diversity fauna of *Ambocoelia* and chonetids. The *Stereolasma* shale can be traced eastward to Knights Falls (Locality 14) and is an excellent mappable unit, the top of which is a well defined upper Mottville contact. Limestone "a" at Locality 5 is a very impure, thin bedded dark gray, crinoidal limestone.

At Jacknife Ravine (Locality 6) 4 feet (1.2 m) of Mottville are exposed: 1 foot (0.3 m) of crinoidal limestone "a" containing large numbers of the brachiopod *Mediospirifer* at its base, overlain by 1.5 feet (0.5 m) of calcareous shale, a second 6 inch (15 cm) lenticular limestone (hereafter referred to as limestone "b") and 1.5 feet (0.5 m) of *Stereolasma* shales. Limestone "a" is more diverse here than at Locality 5, for in addition to *Mediospirifer,* it contains chonetids, and the spiriferid brachiopods *Nucleospira concinna* and *Mucrospirifer.* The *Stereolasma* shales also yield a more diverse assemblage including the rugose coral *Amplexiphyllum hamiltoniae* and the brachiopods *Rhipidomella vanuxemi* and *Douvillina*. *Ancyrocrinus* anchors are common, as are gastropods and *Ambocoelia*.

At the western margin of the Tully 15 minute quadrangle, south of Cedarvale (Locality 7), just over 4 feet (1.2 m) of Mottville is exposed, containing two well defined units, inserted between limestone "a" and the *Stereolasma* shales. All of the contacts here are gradational. Limestone "a" is 1 foot (0.3 m) of dark crinoidal limestone containing a faunal assemblage similar to that at Jacknife Ravine with the addition of the large rugose coral *Heliophyllum halli*. Immediately above is 1.5 feet (0.5 m) of dark gray, calcareous, fossiliferous shale characterized by *Tropidoleptus carinatus, Mucrospirifer,* chonetids and *Bembexia,* which in turn is overlain by crinoidal limestone "b" that is 8 inches (20 cm) thick. The *Stereolasma* shales rest directly on limestone "b" and are 1 foot (0.3 m) thick. Limestone "b" lithologically resembles limestone "a" and yields a low diversity assemblage of orthocones, the large trilobite *Dipleura*, small brachiopods and crinoid stems.

Throughout the remainder of the Tully Valley region (Localities 8 through 11) limestone "b" abounds with tabulate and rugose corals. This unit was named the Case Hill Coral Bed (Grasso, 1968), for exposures in Case Hill Ravine (Locality 9). The Case Hill Coral Bed is dominated by the solitary rugose coral *Heliophyllum halli,* while *Favosites argus, F. alpenensis* and *F. arbuscula* are the dominant tabulates. *Heterophrentis simplex, Cystiphylloides americanum,* and *Favosites hamiltoniae* are conspicuous members of the fauna. This assemblage is interesting in that it includes favositids, such as *Favosites argus*, previously unknown below the Ludlowville Formation in New York (Grasso, 1968).

Limestone "b" extends eastward to the base of Knights Falls (Locality 14) where it is 2.5 feet (0.7 m) of coarse, coquinite-like, crinoidal limestone with a high diversity assemblage of brachiopods and pectenid bivalves. The diversity of corals has decreased compared to the Tully Valley but the rugose coral *Heterophrentis* is conspicuous near the top of the unit.

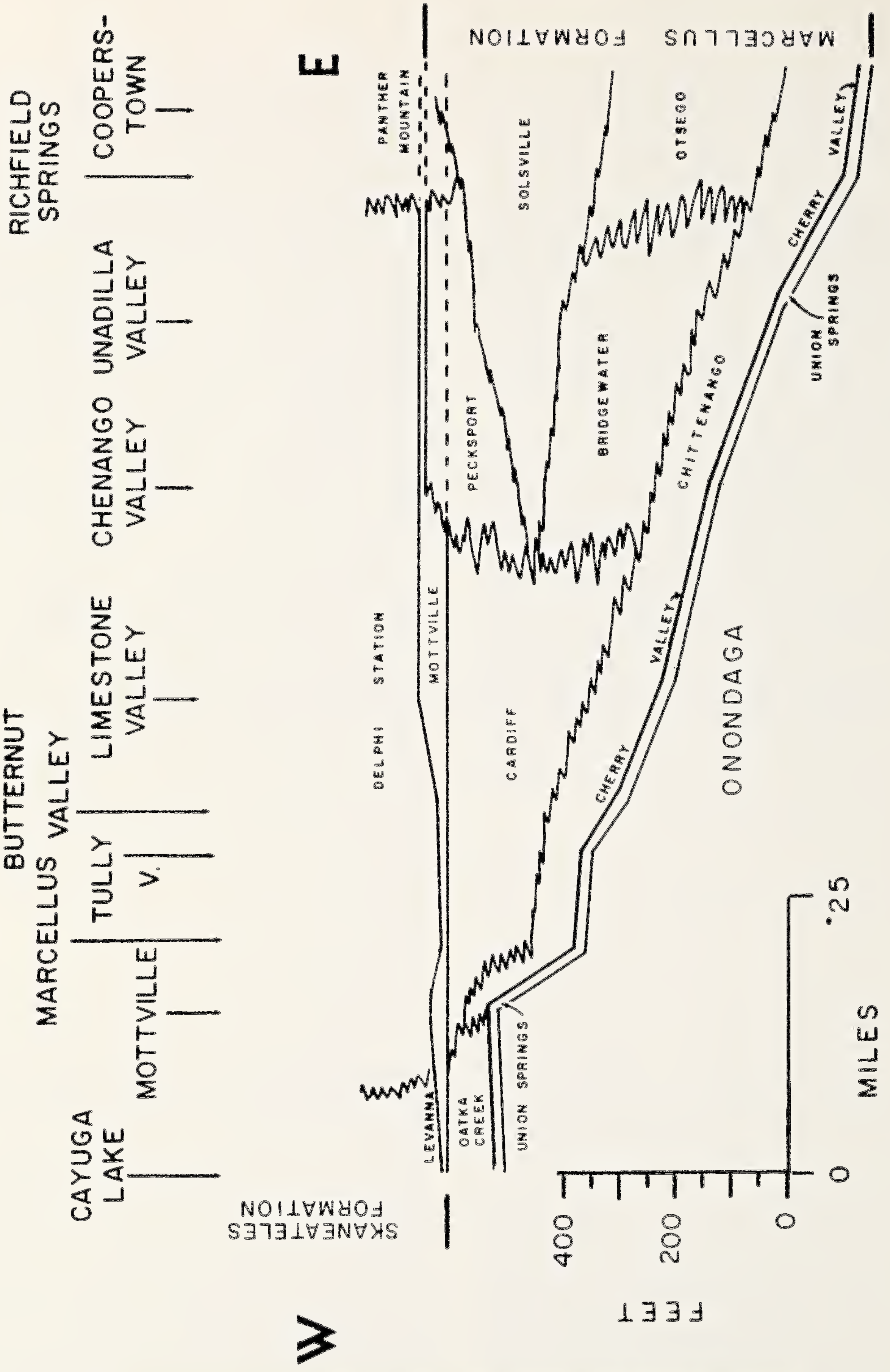


Figure 3. Marcellus—Lower Skaneateles stratigraphic relationships with the redefined Mottville.

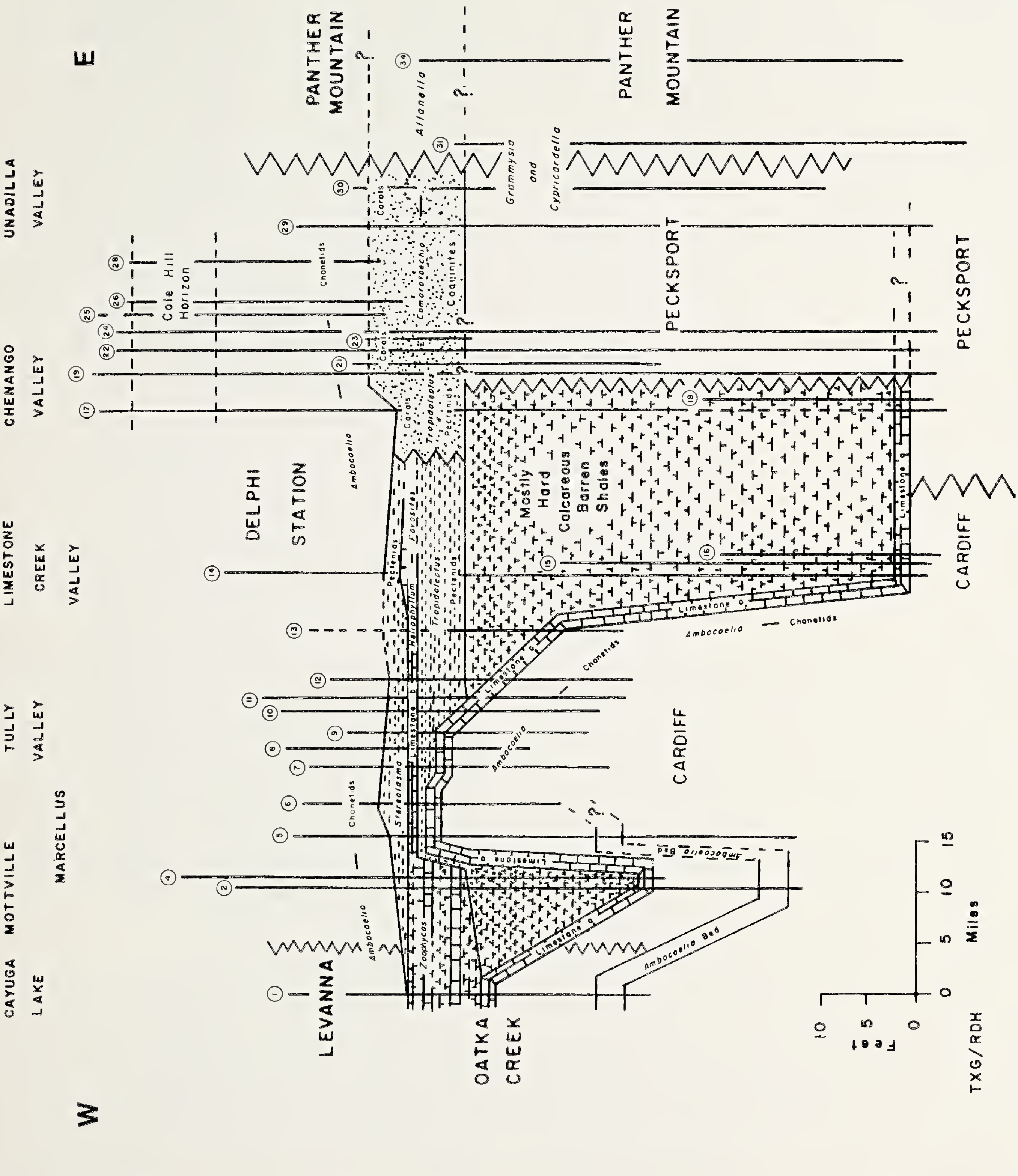


Figure 4. Stratigraphic relationships of the redefined Mottville. Vertical lines represent sections examined. All sections are projected into the vertical plane A-A′ shown on Figure 1. Numbered sections are listed and described in Appendix I.

The *Tropidoleptus*-rich shales, below limestone "b," continue eastward from Locality 7 and thicken to 5 feet (1.5 m) at the base of Pratt's Falls (Locality 13) and to nearly 10 feet (3 m) at Knights Falls. At Knights Falls pectenid bivalves are abundant in these shales. Also, at this locality, the fauna of the *Stereolasma* shale is modified; the *Stereolasma* are smaller in size, less abundant and are associated with more bivalves, especially pectenids, and fenestellid bryozoans.

Between limestone "a" and the *Tropidoleptus* shales is a nearly barren sequence of hard, clinky, calcareous shales that first appear at Pratts Falls (Locality 13) where several feet are poorly exposed. At Knights Falls these barren shales expand to 42 feet (13 m) of strata. The old railroad tunnel northwest of Cazenovia (Locality 15) exposes 36 feet (11 m) of the barren shale interval which caps the section. The barren shale interval rests on 1 to 2 feet (0.5 m) of limestone "a" containing *Mediospirifer*. These nearly barren shales are the hard calcareous shales of Cooper (1930, p. 217). They may correlate with Smith's (1935) hard shales overlying limestone "a" at Mottville, although the Cazenovia barren shales pinch out west of Pratts Falls (Fig. 4).

Limestone "a" can be traced as far as Pine Woods (Locality 18) where it is 8 inches (20 cm) of dark gray, calcareous, silty, mudstone containing some crinoidal material. East of here limestone "a" apparently passes into the upper Pecksport Member. The barren shales grade eastward into the silty shales of the upper Pecksport Member, which in the Unadilla Valley region abound in the bivalves *Grammysia* and *Cypricardella* (Localities 29 and 30, Fig. 4).

There are no exposures of the upper Mottville (*Tropidoleptus* through *Stereolasma* interval) between the Limestone Creek Valley and Chenango Valley, a distance of approximately 15 miles. North of Morrisville Station (Locality 17) this upper Mottville interval commences with 3 feet (1 m) of crossbedded, calcareous, fine sandstone, with coquinites of *"Camarotoechia"* and *Allanella tullius* along with a few *Favosites*, which caps a 25 foot (7.6 m) waterfall. Above the coquinite horizon are 4 feet (1.2 m) of thin bedded, bioturbated, fine sandstone yielding a bivalve assemblage of assorted pectenids, along with *Modiomorpha, Grammysia* and *Goniophora*. The brachiopod *Tropidoleptus carinatus* is common, along with some *Productella*, and possibly *Paraspirifer*. These beds may be a lateral equivalent of the *Stereolasma* shales. Above are the fine grained dark shales of the Delphi Station containing chonetids, *Ambocoelia, Bembexia*, and the inarticulate brachiopod *Craniops*. In the face of the falls, beneath the capping sandstones, hard, calcareous, sparsely fossiliferous shales, containing occasional clusters of *"Camarotoechia"* are exposed for 25 feet (7.6 m). Downstream from the base of the falls the section is covered for approximately 18 feet (5.5 m). Below the covered interval is a small exposure, about 2 feet (0.6 cm) thick, of hard, dark, very calcareous shales containing many small crinoid stems, *Mediospirifer audaculus* and *Mucrospirifer*. This interval is thought to represent a degraded facies of limestone "a." If this correlation is correct the Mottville is approximately 50 feet (15 m) thick at Locality 17. The inferred limestone "a" horizon occurs at an elevation of approximately 1380 feet above sea level, agreeing favorably with the position of limestone "a," at Locality 18, approximately 2 miles east and 1 mile south of Locality 17. Locality 17 represents the easternmost exposure of both the lower and upper contacts of the Mottville as defined to the west.

East of the Chenango Valley only the upper, sandy, coquinite-bearing portion of the Mottville can be recognized. The essential character of the Mottville, as defined above, remains nearly constant across this region to the Unadilla Valley. It is generally 10 to 15 feet (3 to 4.5 m) of bioturbated, ripple-marked, crossbedded, fine, bluish-gray, brown-weathering, calcareous sandstone, containing coquinites of *"Camarotoechia"* and *Allanella tullius*. East of the Unadilla Valley phosphatic pebbles occur sporadically in the unit, sometimes within the coquinites. The brachiopods *Paraspirifer* and *Tropidoleptus*, and bivalves *Ptychopteria* (*Cornellites*) and *Cypricardella* also are found in the sandy Mottville but are subordinate to *"Camarotoechia"* and *Allanella*. Of special significance is the presence of *Favosites, Heliophyllum* and *Heterophrentis* near the top of the Mottville in the Chenango Valley (Localities 17, 20, 21, 22), in the Unadilla Valley (Locality 30) and near Richfield Springs (Localities 31 and 34). The presence of these corals suggests a time equivalence with limestone "b" to the west. East of the Unadilla Valley region the Mottville cannot be readily discerned, but it appears to be present on the west side of Otsego Lake (Localilty 35, Fig. 1).

In the overlying Delphi Station Member, between the Chenango and Unadilla Valleys, there is a 6 to 10 foot (1.8 to 3 m) thick interval with abundant large bivalves and the trilobite *Dipleura*, approximately 30 feet (9 m) above the top of the Mottville. The author refers to this bed as the Cole Hill horizon for exposures at Cole Hill on the west side of the Sangerfield Valley (Locality 25). The Cole Hill horizon extends from Locality 17 near Morrisville Station to the Brookfield Valley (Locality 28) and possibly to Locality 32 southwest of Richfield Springs (Fig. 1). The high diversity assemblage includes the bivalves *Limoptera, Actinopteria, Actinodesma* (*Glyptodesma*), brachiopods, and *Dipleura*.

MOTTVILLE FAUNA AND COMMUNITIES

The Mottville Member possesses a low diversity fauna at Cayuga Lake that increases rapidly east of Marcellus to a maximum in the Tully Valley (Fig .5). Some of this rise reflects an increase in diversity within limestone "a" but most is the result of the eastward appearance of limestone "b," and the *Tropidoleptus* and *Stereolasma* shales below and above this limestone.

In the Tully Valley, limestone "a" and the *Tropidoleptus* shales account for most of the species diversity, but eastward in the Limestone Creek region limestone "a" diversity has decreased dramatically. The total Mottville diversity here is

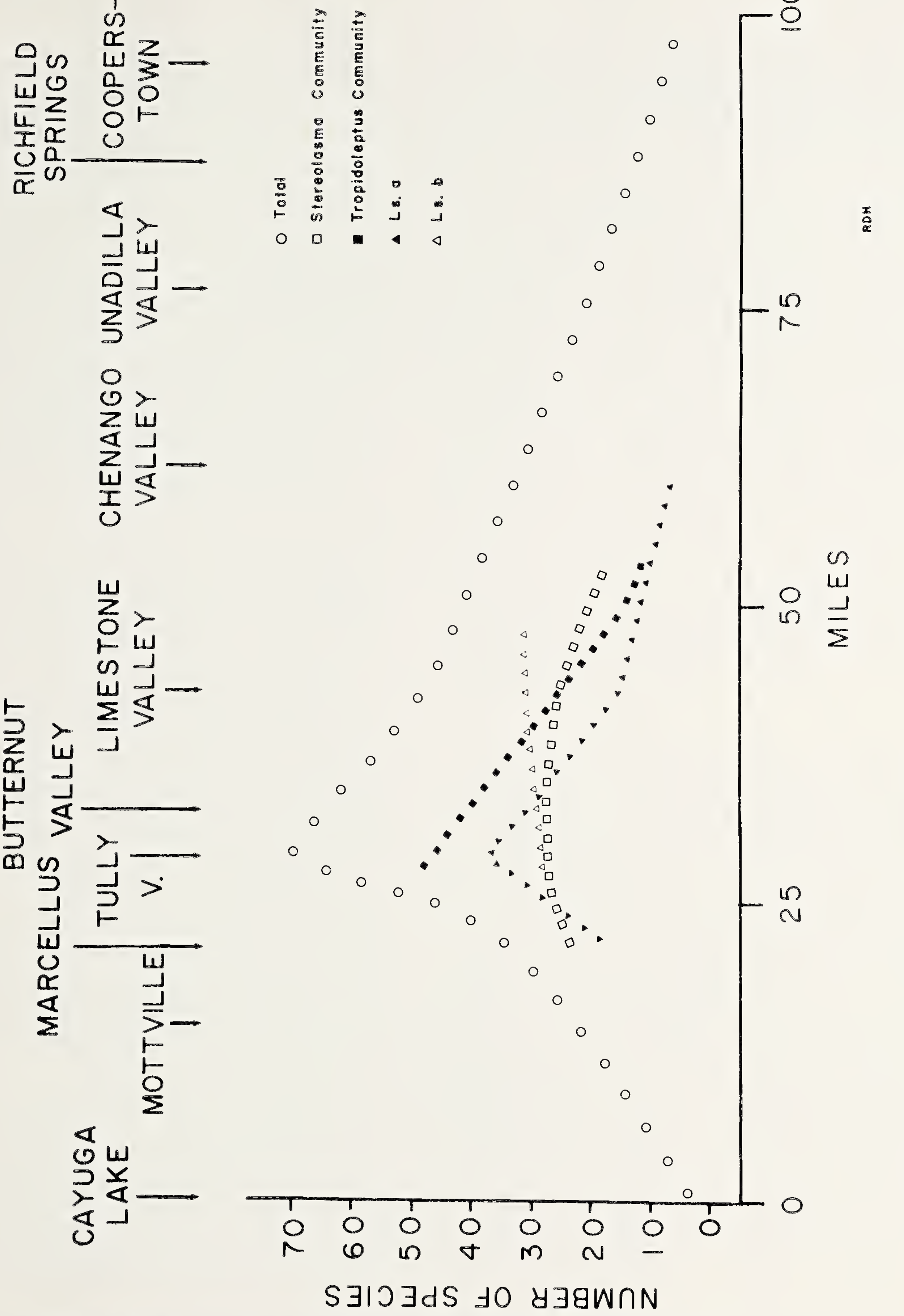


Figure 5. Mottville faunal diversity curves. For each meridian the number of species shown on any curve represents the total species summed from many localities along and adjacent to that meridian. Faunal data are listed in Appendix II.

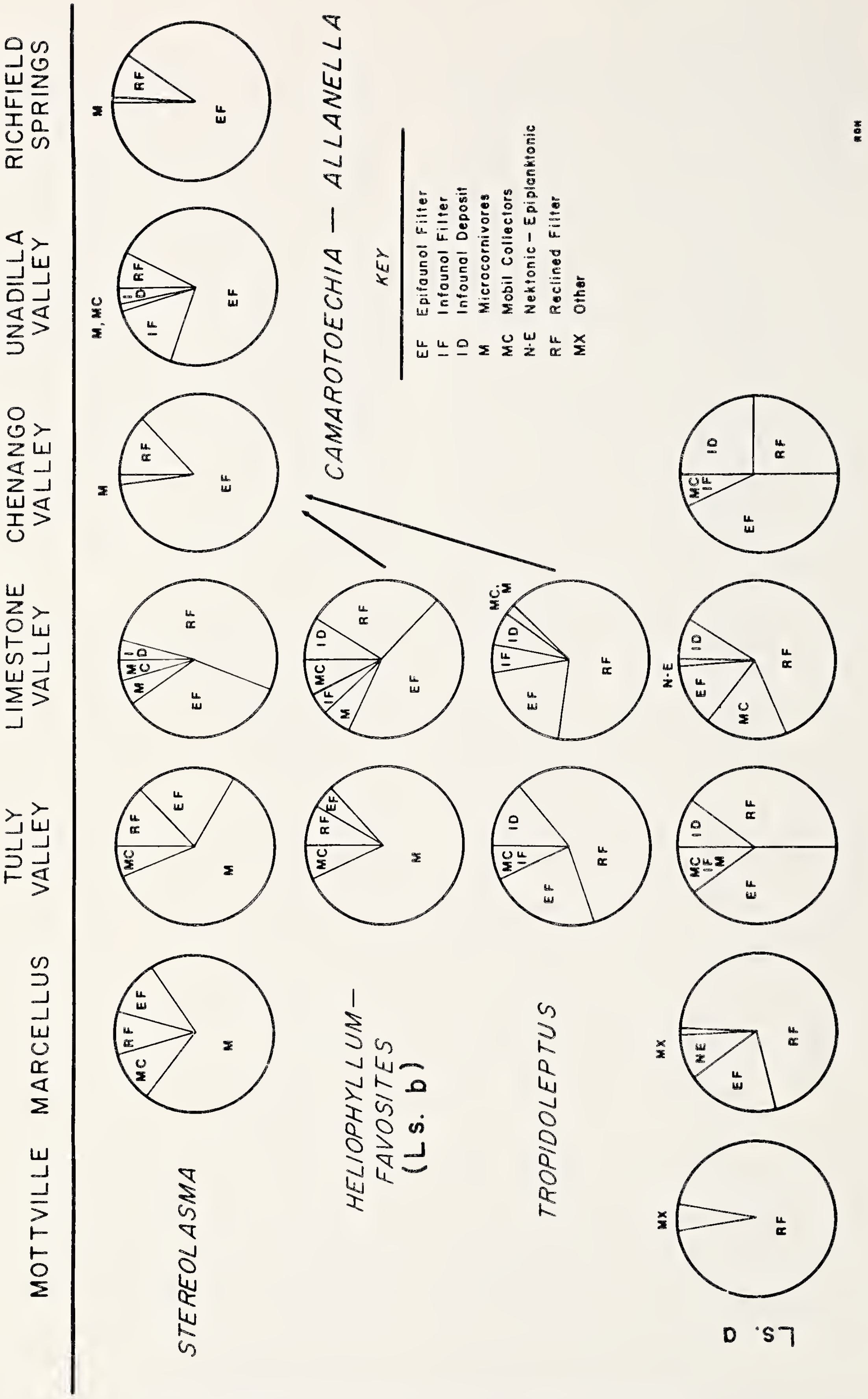


Figure 6. Mottville trophic structure. Faunal data are listed in Appendix II.

shared nearly equally by the *Tropidoleptus* shales, limestone "b", and the *Stereolasma* shales. The decrease in diversity east of Limestone Creek Valley reflects the faunas of the sandy Mottville, that are dominated by coquinites of *"Camarotoechia"* and *Allanella*. The occurrence and abundance data for the Mottville faunas are tabulated in Appendix II.

Richard Liebe of the State University of New York College at Brockport processed several samples for conodonts. The author identified the shallow water species *Icriodus latericrescens latericrescens* from limestone "a" near Mottville (Locality 3) and the Tully Valley (Locality 11). Limestone "b" yielded the same species from Locality 7. A partial specimen, questionably assigned to *I. latericrescens latericrescens* was recovered from the sandy Mottville east of Hamilton (Locality 21). These records confirm earlier reports of *I. latericrescens latericrescens* in the Mottville. The Mottville and succeeding Levanna and Delphi Station Members have been tentatively assigned to the *ensensis* conodont zone (Klapper, 1981, p. 61-62).

Limestone "a" Fauna

The fauna of limestone "a" is heterogenous, although from Jacknife Ravine (Locality 6) to Locality 18 in the Chenango Valley, *Mediospirifer audaculus* is more or less consistently abundant throughout its lower part.

At Mottville, reclining filter feeders such as chonetids, *Ambocoelia*, and *Longispina*, along with *Mucrospirifer* and *Schuchertella*, dominate the moderately diverse assemblage (Fig. 6). The auloporid coral *Aulocystis* is found as well as trilobites (*Phacops, Greenops and Dipleura*) and gastropods (*Platyceras, Palaeozygopleura* and *Bembexia*). From Jacknife Ravine to the Tully Valley at Bear Mountain Ravine and Rainbow Creek (Localities 10 and 11) diversity increases with the corals *Heliophyllum, Heterophrentis* and *Favosites* reaching their peak abundances in this unit. In the Tully Valley, the fauna of limestone "a" resembles a lower grade limestone "b" fauna. Here epi- and endobyssate filter feeding bivalves are particularly common including *Pterinopecten, Ptychopteria* (*Cornellites*), *Leiopteria, Glyptodesma*, and *Mytilarca*. The infaunal filter feeding bivalves *Gosselettia* and *Modiomorpha* are also well represented. These biofacies changes across the Tully Valley may be indicative of shallowing conditions to the southeast.

In Limestone Creek Valley and near Cazenovia the limestone "a" fauna resembles that of the Marcellus-Mottville region. East of Morrisville (Localities 17 and 18) limestone "a" is poorly exposed and contains *Mediospirifer, Mucrospirifer*, chonetids and the infaunal deposit feeding bivalves *Nuculoidea* and *Palaeoneilo*. Limestone "a" everywhere overlies dark shales that generally possess a low diversity assemblage of *Ambocoelia*, chonetids, *Leiorhynchus* and *Pterochaenia fragilis*. However, in the Tully Valley (Locality 10) and the Butternut Valley (Locality 12) the small, discoid, rugose coral *Microcyclus* was found in abundance, with several other species, 14 feet (4 m) and 12 feet (3.6 m) respectively, below limestone "a."

There is no clearly defined community in limestone "a" across its outcrop belt, although the author (Grasso, 1983) earlier referred its fauna to the *Mediospirifer* Community. The unit is sometimes poorly exposed and even when it is well exposed it forms the crests of nearly vertical waterfalls making sampling difficult. The data that have been gathered suggest the existence of at least three biotopes during limestone "a" time: (1) a low diversity assemblage dominated by reclining filter feeding brachiopods and *Mediospirifer audaculus* present near Mottville (Localities 3 and 4) and reappearing in the northern Limestone Creek Valley at Pratts Falls (Locality 13) and northwest of Cazenovia (Locality 15); (2) a slightly higher diversity assemblage of epifaunal, filter feeding, pedunculate brachiopods such as *Nucleospira, Rhipidomella, Mediospirifer*, and rugose and favositid corals characteristic of the northern Tully Valley (Localities 7, 8, and 9), southern Limestone Creek Valley (Locality 14) and northern Chenango Valley (Locality 17); and (3) an assemblage present only in the southern Tully Valley (Localities 10 and 11) containing *Mediospirifer audaculus*, pectenid and infaunal filter feeding bivalves, and a few rugose corals.

Tropidoleptus Community

The *Tropidoleptus* Community underlies limestone "b" between Locality 7 at the west edge of the Tully quadrangle to Knights Falls (Locality 14) in the Limestone Creek Valley, a distance of approximately 20 miles. Diversity is highest in the Tully Valley and decreases eastward (Figure 5). The community is dominated by the reclining orthid *Tropidoleptus carinatus*. Other reclining, filter feeding, brachiopods include chonetids, *Ambocoelia, Mucrospirifer, Protoleptostrophia* and *Pholidostrophia nacrea*. Epifaunal attached brachiopods are represented by *Mediospirifer audaculus, Nucleospira*, and *Elytha fimbriata*; bivalves by *Actinopteria, Leiopteria, Pterinopecten, Ptychopteria* (*Cornellites*), *Modiomorpha, Macrodon, Nyassa*, and some unidentified pectenids. The bivalves are more abundant in the Limestone Creek Valley at Knights Falls than farther west (Appendix II). The deposit feeding bivalves *Nuculoidea, Nuculites*, and *Paleoneilo* are minor constituents of the community; the trilobites *Phacops, Greenops* and *Dipleura* and the gastropods *Bembexia, Palaeozygopleura* and *Bellerophon* are present, though not abundant. At its westernmost exposure (Locality 7) the *Tropidoleptus* Community is typically less diverse and faunally resembles that of the underlying limestone "a."

Heliophyllum-Favosites Community

Confined to the limestone "b," this community is dominated by microcarnivorous solitary rugose and tabulate corals. It extends from Locality 7 to Knights Falls. The rugose corals present, in decreasing abundance, are *Heliophyllum halli, Heterophrentis* and *Cystiphylloides. Favosites argus, F. alpenensis* and *F. arbuscula* are the dominate tabulates; *F. hamiltoniae* is present but not abundant. The tabulates are most abundant in the Tully Valley (Localities 8 through 11). It is a moderate diversity community that includes brachiopods (*Rhipidomella, Nucleospira*), bivalves (*Actinopteria*), bryozoans (*Fenestella*), trilobites (*Phacops, Dipleura*), gastropods (*Palaeozygopleura, Bembexia, Mourlonia*) and *Zoophycos* spreiten.

In the Limestone Creek Valley,. the composition of the *Heliophyllum-Favosites* Community is modified from that in the Tully Valley (Fig. 6). The rugose corals are represented by *Heterophrentis* (common) and an occasional *Heliophyllum*; the favositids are absent. In addition, sediment tolerant taxa are more abundant than farther west and a few new species, not present to the west, make their appearance. New taxa include the bivalves *Ptychopteria* (*Cornellites*) and *Pterinopecten* and the reclining brachiopods *Tropidoleptus* and *Pholidostrophia* (Fig. 6).

Stereolasma Community

The uppermost part of the Mottville from Falls Ravine (Locality 5) to Knights Falls is characterized by the *Stereolasma* Community. It is moderately diverse (Fig. 5) and dominated by the small rugose coral *Stereolasma rectum* along with *Amplexiphyllum* in the westernmost localities. East of the Tully Valley *Amplexiphyllum* is not present although *Stereolasma* continues to Knights Falls, albeit somewhat reduced in abundance. The pedunculate brachiopod *Rhipidomella* is consistently abundant in this community while *Nucleospira* is less so. Chonetids, *Ambocoelia* and *Pholidostrophia* are conspicuous along with the trilobite *Phacops* and gastropods such as *Platyceras, Bembexia* and *Mourlonia*.

In the Limestone Creek Valley, the *Stereolasma* Community shows an increase in proportion of epifaunal filter feeders in much the same way as the underlying *Heliophyllum-Favosites* and *Tropidoleptus* Communities discussed above. Pectenid bivalves such as *Actinopteria* increase in abundance, as do fenestellid bryozoans. The infaunal filter feeding bivalves *Nyassa, Cypricardella* and *Paracyclas* also are present, as well as deposit feeding nuculid clams (Fig. 6).

"Camarotoechia"-Allanella Community

The *"Camarotoechia"-Allanella* Community, associated with the sandy facies, extends from the west side of the Chenango Valley (Locality 17) to the east side of the Unadilla Valley (Locality 30) and comprises all of the recognizable Mottville in this region. At its westernmost exposure it contains abundant *Tropidoleptus* and pectenids which suggests time-equivalence with the *Tropidoleptus* through *Stereolasma* Communities of the Limestone Creek Valley. Localized abundance of rugose (*Heterophrentis, Heliophyllum*) and tabulate corals (*Favosites*) at Localities 17, 20, 21, 22, 30, 31, and 34 also support this conclusion. The eastward continuation of the *"Camarotoechia"-Allanella* Community is uncertain, although it may extend to Otsego Lake (Locality 35).

The Community is dominated by attached epifaunal, filter feeding taxa, primarily brachiopods, such as the rhynchonellid *"Camarotoechia"* and the spiriferid *Allanella tullius* (=*Adolphia?*)which mainly occur in lag concentrated coquinites (Fig. 6). *Mucrospirifer* also is found in the coquinites but these individuals may have been pedicle attached (J. Brower, Syracuse University, personal communication). *Paraspirifer* is a moderately important element of the fauna, especially in the Unadilla Valley and more eastern localities. The reclining brachiopods *Productella, Tropidoleptus,* and *Protoleptostrophia* are represented in this community at many localities. Epibyssate and endobyssate bivalves are common, e.g., *Ptychopteria* (*Cornellites*), and *Mytilarca*, along with smaller numbers of *Goniophora*. The infaunal filter feeding bivalves *Cypricardella, Modiomorpha,* and *Grammysia* occur sporadically throughout the community. The only trilobite found is *Dipleura*, gastropods are only rarely present, and *Zoophycos* traces are ubiquitous.

The *"Camarotoechia"-Allanella* Community appears to be very similar to the *Cyrtospirifer-Camarotoechia* Community of McGhee (1976) and McGhee and Sutton (1981) in the Late Devonian of the central Appalachians, as well as Baird and Brett's (1983) *Allanella* (shallow shelf: variable turbidity-high energy) association in the Ludlowville Formation (Hamilton Group) of west-central New York State.

PALEOENVIRONMENTS

The Oatka Creek, Cardiff and Pecksport Members of the Marcellus Formation are characterized by a mixed, low diversity assemblage of the benthonic brachiopods *Ambocoelia*, chonetids, *Leiorhynchus* and the thin shelled bivalve *Pterochaenia fragilis* in a fine grained, soft, dark gray to black shale. The benthic association is dominated by small reclined epifaunal brachiopods without a functional pedicle in the adult stage. The deeply incurved pedicle valves of *Ambocoelia* and chonetids make them ideally suited for life on soft substrates. *Leiorhynchus* may have lived attached to plants and shells in clusters as suggested by Bowen and others (1974). The benthic and lithic characteristics suggest an environment of relatively deep, low energy, possibly poorly oxygenated or dysaerobic water. The Levanna and lower Delphi Station Members of the Skaneateles Formation probably were also deposited under similar environmental conditions (Fig. 4). Baird and Brett's (1983) *Leiorhynchus* (basin center) and *Ambocoelia*-nuculid (basin slope-outer shelf) associations are applicable to the shales below and above the Mottville.

The coarse crinoidal lithology and relatively higher faunal diversity of limestone "a" suggests generally shallower water and/or slower deposition than that which prevailed in late Marcellus time. The overall shallowing scenario for limestone "a" is complicated by an inferred regional high with a southeastward upslope gradient in the Tully Valley region. An analogous situation has been suggested by Baird and Brett (1983) for the Ludlowville Formation in west-central New York State. The increase in faunal diversity, especially in corals, and a slightly coarser encrinitic lithology southeastward across the Tully Valley lends credence to this shallow water hypothesis.

The general regressive nature of the lower portion of limestone "a" is not pronounced west of the Tully Valley, where deposition was in a deeper, basinward setting. Also, bioturbation may have obscured the natural faunal associations. In the Chenango Valley region, similar bathymetric conditions may have prevailed which, coupled with greater clastic influx, caused limestone "a" to be indistinguishable from the upper Pecksport. Another possibility is that limestone "a" represents a complete minor regressive cycle in the Tully Valley, bracketed by slightly deeper water brachiopods such as *Mediospirifer* below and *Tropidoleptus* above, but only a hemi-cycle eastward, prematurely terminated by the rapid deposition of the nearly barren shales above.

The nearly barren shale interval is thought to represent rapid sedimentation, probably from the southeast, into a quiet water, deeper, environmental setting. This shallowing upward facies grades rapidly into the overlying *Tropidoleptus* Community or its correlative, the lower portion of the *"Camarotoechia"-Allanella* Community. The upper part of the barren shale interval grades eastward into the silty shales of the upper Pecksport Member which abound in *Grammysia* and *Cypricardella*.

The high diversity *Tropidoleptus* Community and the overlying *Heliophyllum-Favosites* and *Stereolasma* Communities record the culmination of the shallowing upward, regressive depositional cycle, with maximum regression or "core" of the cycle represented by the *Heliophyllum-Favosites* Community in limestone "b". This hypothesis is supported by the coarse crinoidal texture and high diversity assemblage characterized by large corals of limestone "b". The *Stereolasma* Community represents a return to slightly deeper water conditions (transgression), similar in bathymetry to the *Tropidoleptus* Community below but with lower turbidity allowing *Stereolasma* to become widespread. These conditions are very similar to those described by Gray (1983) for the Centerfield Member. The *Tropidoleptus* and *Stereolasma* Communities are analogous to Baird and Brett's (1983) upper slope-shelf associations of *Athyris-Pseudoatrypa* (low-moderate turbidity) and/or *Tropidoleptus-Zoophycos* (moderate-high turbidity). The *Heliophyllum-Favosites* Community equates with their *Heliophyllum-Pentamerella* (shelf): low-moderate turbidity) association.

In the east the *"Camarotoechia"-Allanella* Community inhabited an offshore sandy shelf or bar environmental setting. It occupied an area of high energy, shallow water, and variable turbidity as shown by the lithology, sedimentary structures (ripple marks and cross-bedding), coquinites and occasional corals. The coquinites probably represent lag concentrates that were not transported great distances. Although many brachiopod valves are separated, most are whole and display little abrasion. In addition, the coquinites are found at various horizons within the sandy Mottville but generally more so in the lower or middle part and only occasionally at or near the upper contact. At a few localities they are poorly developed or lacking, while at others more than one coquinite horizon is present. Phosphate pebbles can be found within the coquinites (Localities 30 and 31). Thus, the shallowing upward cycle is not symmetrical in the eastern portion of the study area. It is further complicated by local discontinuities, probably due to lateral shifting of sediment supply and the vagaries of current intensity and direction. Perhaps minicycles or portions of minicycles operating within the overall regressive theme further obscure the relationships. In general, the shallowest part of the cycle (peak regression) would correspond to the sandy conquinites, especially at those localities where they occur below the upper contact and lack phosphate nodules. The uppermost fine grained sandy or silty Mottville then would correlate with the *Stereolasma* Community and would indicate the start of a transgressive phase.

The underlying *Grammysia* and *Cypricardella* silty shales of the upper Pecksport are intermediate in depth from the deeper water lower Pecksport below and the shallower water Mottville above. Grasso and Wolff (1977) assigned this interval of the Pecksport to the outer delta platform realm. The uppermost part of these beds may in fact correlate with part of the *Tropidoleptus* beds farther west.

The convergence of limestones "a" and "b" westward is interpreted as a condensed sequence, with limestone "b," the underlying litho- and biofacies, and the *Stereolasma*-bearing beds nearly merging with limestone "a" at Jacknife Ravine in the Marcellus Quadrangle. This hypothesis is suggested because of the gradational contacts of the various Mottville facies vertically at any one locality and the absence of any indication of a disconformity above limestone "a." West of the Marcellus meridian, limestones "a" and "b" diverge abruptly so that at Mottville, and localities just to the west, they are separated by 15 to 20 feet (4.5 to 6 m) of nearly barren, hard, calcareous shales. Furthermore, these shales grade upward into 5 to 8 feet (1.5 to 2.5 m) of finely crystalline, hard limey shales with *Zoophycos* swirls that probably are equivalent to the eastward synjacent *Tropidoleptus* and *Heliophyllum-Favosites* (limestone "b") Communities and perhaps a portion of the *Stereolasma* Community as well (Fig. 7). Further westward, at Cayuga Lake, limestone "a" and the *Zoophycos*-bearing mudstones (swirl-rock) completely converge pinching out the barren shale interval. The *Zoophycos* swirl-rock (limestone "b"

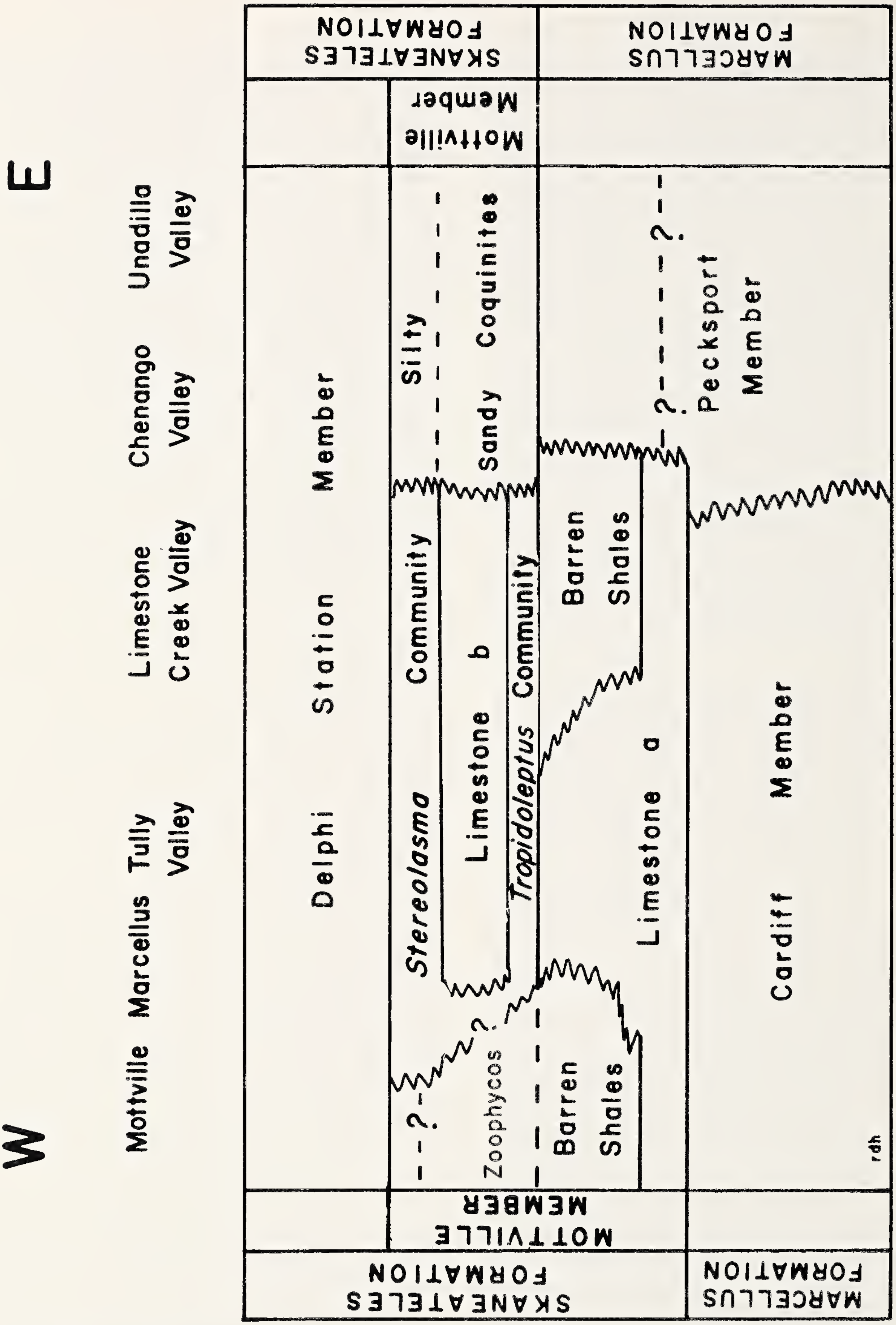


Figure 7. Summary of redefined Mottville relationships. See text for details. Figure is schematic; units not shown to true scale.

equivalent) represents the shallowest part of the cycle in the westernmost portion of the study area and is equivalent to the Stafford Limestone west of Cayuga Lake.

The Levanna and lower Delphi Station Members, overlying the Mottville, are basinal deposits similar to the Cardiff (although the lower Delphi Station may not be quite as deep as the upper Cardiff) and represent a return to deeper water, following the transgression, that began with the *Stereolasma* Community.

The Cole Hill horizon of the lower Delphi Station Member in the eastern portion of the study area carries a distinctive fauna called the *Limopteria* Community by Grasso and Wolff (1977). Large pectenid bivalves, pedunculate brachiopods and trilobites dominate the assemblage. Other constituent taxa include gastropods and *Zoophycos*. The fauna is dominated by filter feeders (nearly 75 percent) while mobile epifaunal taxa such as gastropods and trilobites account for the remainder. The bivalves include the epibyssate and endobyssate forms *Limopteria, Mytilarca, Ptychopteria* (*Cornellites*), *Pseudaviculopecten, Actinodesma* (*Glyptodesma*), *Modiomorpha, Goniophora* and *Cimitaria*. *Grammysia* and *Cypricardella*, the infaunal filter feeding bivalves, also are abundant. *"Camarotoechia," Mucrospirifer, Paraspirifer, Mediospirifer*, and *Tropidoleptus* are the dominant brachiopods. The trilobites *Dipleura* and *Greenops* are abundant while the gastropods *Bembexia* and *Palaeozygopleura* are common and conularids infrequent. The assemblage is characterized by fixed and free epifaunal filter feeders and mobile and fixed infaunal filter feeding bivalves. Vagrant benthonics are strikingly conspicuous. Lithologically the Cole Hill Horizon is a calcareous, arenaceous and silty shale or subgraywacke, containing, at some localities, low angle planar crossbedding (Grasso and Wolff, 1977). The inferred environment of the *Limopteria* Community is a middle delta platform or shelf where current activity was moderate to high. Normal marine conditions prevailed and the substrate was stable and firm. Sufficient organic material was present to support the varied adaptive types described above.

Below the Cole Hill horizon but above the Mottville is a low to moderate diversity fauna containing *Ambocoelia*, chonetids and nuculid bivalves along with other bivalves and brachiopods in a dark, splintery, silty shale matrix. This assemblage and lithology is gradually replaced upward by the *Limopteria* Community in coarser grained clastics. Therefore, the post- Mottville change back to deeper water occurred in the east but is not as striking as in the west. Selleck (1983) suggests that the highest diversity assemblages of the Hamilton Group are diagnostic of intermediate water depths. Therefore, the Cole Hill horizon probably represents the start of another, smaller, regressive cycle or hemicycle probably reaching only an intermediate water depth.

SUMMARY AND CONCLUSIONS

The Mottville Member is redefined as a mappable unit from Cayuga Lake to the east side of the Unadilla Valley, south of West Winfield. The 8 feet (2.5 m) of Mottville at Cayuga Lake (limestone "a" and *Zoophycos* swirl-rock) is equivalent to the 25 foot (7.5 m) section at Mottville, beginning with limestone "a" and succeeded by barren shales capped by *Zoophycos* swirl-rock. The Mottville succession is subdivided into five mappable litho- and biofacies (communities) in the region between the Tully and Limestone Creek Valleys that total approximately 50 feet (15 m) of section (Fig. 7). These are, in ascending order: (1) limestone "a"; (2) nearly barren, hard, calcareous shales; (3) *Tropidoleptus* Community; (4) limestone "b"-*Heliophyllum-Favosites* Community and (5) *Stereolasma* Community. The upper three units pass eastward into a 10 to 20 foot (3 to 6 m) thick sandy, coquinite bearing facies with *"Camarotoechia"* and *Allanella*. Across the same area the lower Mottville of the west grades into the upper Pecksport containing *Grammysia* and *Cypricardella* in its uppermost part.

The vertical succession of lithologies and communities suggests an upward shallowing sequence with peak regression in limestone "b" and the *Zoophycos* mudstones in the west and in the sandy coquinites in the east, although eastern sections have a more complex history. The entire Mottville interval lies between the deeper water, fine dark shales of the Oatka Creek, Cardiff and Pecksport Members (Marcellus Formation) below and the Levanna and Delphi Station Members (Skaneateles Formation) above, all carrying a low diversity mixed assemblage of *Ambocoelia*, chonetids, *Leiorhynchus* and nuculid bivalves. Except for limestone "a," which is a puzzling unit, the eastward lateral litho- and biofacies changes suggest shoaling conditions as well. The Cole Hill Horizon of large bivalves in the lower Delphi Station to the east represents the start of another regressive cycle or subcycle reaching intermediate water depths. The deeper water phase of this regression, just above the Mottville, is not as dramatic as it is to the west. The available lithic and faunal data indicate that the eastward equivalent of the Mottville in the Catskill redbed facies (if it can be recognized at all) should occur at or near the top of a regressive hemicycle and not in a transgressive tongue, as has previously been suggested.

ACKNOWLEDGMENTS

I would like to thank Richard Hamell (Monroe Community College) for assisting me with the field work, drafting all the figures, and reviewing the manuscript. John Cottrell (Monroe Community College and The University of Rochester) also read the manuscript, offered many helpful suggestions and discussed with the author many of the topics contained herein.

James Brower (Syracuse University) and William Kirchgasser (SUNY-Potsdam) critically reviewed the final manuscript. Richard Liebe (SUNY-Brockport) processed many samples for conodonts. I am deeply indebted to Carlton Brett (University of Rochester) for encouraging me to present a preliminary version of this paper at the Geological Society of America Northeast Sectional Meeting at Kiamesha Lake and for the many fruitful discussions we had in and out of the field. Acknowledgement is made to the Research Foundation of State University of New York for Grant-in-Aid Awards 081-7101A and 081-7103A which helped defray the cost of field work and other aspects of the project.

REFERENCES CITED

Baird, C.G. and Brett, C.E. 1983. Middle Devonian detrital shelf-to basin ramp: Ludlowville lithofacies and faunal associations in west-central New York State. Geol. Soc. Amer. Abstr. with Programs 15(3):170.

Bowen, Z.P., Rhoads, D.C., and McAlester, A.L. 1974. Marine benthic communities of the Upper Devonian of New York. Lethaia 7:93-120.

Brett, C.E., Baird, G.C., and Kloc, G.J. 1983. Paleoenvironmental models of the Hamilton Group: cyclic and episodic deposition in a stratified basin. Geol. Soc. Amer. Abstr. with Programs 15(3):172.

_________ and Baird, G.C. 1982. Upper Moscow-Genesee stratigraphic relationships in western New York. New York State Geol. Assn. Guidebook, 54th Ann. Mtg., SUNY at Buffalo, p.19-63.

Cooper, G.A. 1930. Stratigraphy of the Hamilton Group of New York, parts 1 and 2. Amer. J. Sci. 19:116-134, 214-236.

_________ 1933. Stratigraphy of the Hamilton Group, eastern New York, part 1. Amer. J. Sci. 26:537-551.

Fernow, L.R. 1961. Paleoecology of the Middle Devonian Hamilton Group in the Cayuga Lake region. Doctoral dissertation, Cornell University, 209 p.

Grasso, T.X. 1968. New coral bed in the Hamilton Group (Middle Devonian) of central New York. J. Paleont. 42(1):84-87.

_________ 1983. Stratigraphy and depositional environments of the Mottville Member in central and eastern New York. Geol. Soc. Amer. Abstr. with Programs 15(3):170.

_________ and Wolff, M.P. 1977. Paleoenvironments of the Marcellus and lower Skaneateles Formations of the Otsego County region. New York State Geol. Assn. Guidebook, 49th Ann. Mtg., SUNY College at Oneonta, A3 p. 1-50.

Gray, L.M. 1983. Biofacies and depositional environments of the Centerfield "Member" in the northern Appalachian Basin. Geol. Soc. Amer. Abstr. with Programs 15(3):170.

Hall, J. 1843. Geology of New York, Part 4, comprising the survey of the Fourth Geological District. Albany, New York, 525 p.

Klapper, G. 1981. Review of New York Devonian conodont biostratigraphy. *In* Oliver, W.A., Jr. and Klapper, G., eds., Devonian Biostratigraphy of New York, Subcomm. on Devonian Strat., Part 1, p. 57-66.

McGhee, G.R., Jr. 1976. Late Devonian benthic marine communities of the central Appalachian Allegheny Front. Lethaia 9:111-136.

_________ and Sutton, R.G. 1981. Late Devonian marine ecology and zoogeography of the central Appalachians and New York. Lethaia 14:27-43.

Rickard, L.V. 1975. Correlation of the Silurian and Devonian rocks in New York State. New York State Mus. Map and Chart Series No.24, 16 p.

Selleck, B.W. 1983. Cyclic shallowing-upward sequences in the Hamilton Group of central New York. Geol. Soc. Amer. Abstr. with Programs 15(3):170.

Smith, B. 1916. Structural relations of some Devonian shales in central New York. Acad. Nat. Sci. Phila. Proc. 67:561-569.

_________ 1935. Geology and mineral resources of the Skaneateles quadrangle. New York State Mus. Bull. 300, 120 p.

Vanuxem, L. 1842. Geology of New York, Part 3, comprising the survey of the Third Geological District. Albany, New York, 306 p.

APPENDIX I
LOCALITY REGISTER AND UNITS EXPOSED

All quadrangles 7 1/2 minute, elevations approximate, all sections measured with a hand level. Thickness figures include covered intervals, unless noted otherwise. Thicknesses of oldest units represent exposed thicknesses only, at each locality. Thickness and elevation figures given in feet.

1. Union Springs Quadrangle, Cayuga Co. — Great Gully, up from N.Y. Route 90, about 2.3 miles south of Union Springs just south of, and parallel to, Great Gully Road.

Mottville	
Zoophycos swirl-rock	8.0′ Elevation 430′
limestone "a"	0.5′
Oatka Creek	12.5′

2. Skaneateles Quadrangle, Cayuga Co. — Unnamed ravine extending west onto Auburn Quadrangle, up from Parcel Road about 0.4 miles south of its intersection with Miller Road.

Mottville	
Zoophycos swirl-rock	5′
hard calcareous shales	20′
limestone "a"	1.5′ Elevation 810′
Cardiff	20′
Chittenango	5′ (only upper 5′exposed)
Cherry Valley	covered
Union Springs	covered
Onondaga	Elevation top 740′

3. Skaneateles Quadrangle, Cayuga Co. — Small unnamed ravine up from Miller Road, parallel to and 300 feet west of Onondaga-Cayuga County Line Road.

Mottville	
hard calcareous shales	20′
limestone "a"	1.25′ Elevation 820′
Cardiff	30′
Chittenango	29′

4. Skaneateles Quadrangle, Onondaga Co. — Skaneateles Creek, 0.2 miles south of Mottville along the abandoned Skaneateles Short Line Railroad just south of the point where the railroad crosses to the west side of the creek. Type section of Mottville Member.

Mottville	
Zoophycos swirl-rock	5′
hard calcareous shales	20′
limestone "a"	1′ Elevation 820′
Cardiff	10′

5. Skaneateles Quadrangle, Onondaga Co. — Falls (Thorne) Ravine on the southwest side of Guppy Gulf parallel to and 600 feet east of Gully Road just north of its intersection with the New Seneca Turnpike.

Delphi Station	Not measured
Mottville	
Stereolasma Community	2′
limestone "a" and limestone "b"?	1.5′ Elevation 830′
Cardiff	68′

6. Marcellus Quadrangle, Onondaga Co. — Jacknife Ravine up (south) from Pleasant Valley Road, about 1 mile east of its intersection with N.Y. Route 175 and east of Slate Hill Road.

Delphi Station	Not measured
Mottville	
Stereolasma Community	1′
limestone "b"	6′
limestone "a"	1′ Elevation 920′
Cardiff	151′
Chittenango	62′

7. Marcellus-South Onondaga Quadrangles, Onondaga Co. — Unnamed ravine on the southwest side of the Tully Valley, 2 miles south of Cedarvale and 0.7 miles southwest of Nichols Corners. It is the first ravine 0.4 miles south of Tanner Road at the west edge of the South Onondaga Quadrangle.

Delphi Station	Not measured
Mottville	
Stereolasma Community	1′
limestone "b"	8′
Tropidoleptus Community	1.5′
limestone "a"	1′ Elevation 840′
Cardiff	195′
Chittenango	47′

8. South Onondaga Quadrangle, Onondaga Co. — Lords Hill Ravine, south of Ironsides, just east and parallel to N.Y. Route 80.

Delphi Station	Not measured
Mottville	
Stereolasma Community	1′
limestone "b"	9′
Tropidoleptus Community	2.25′
limestone "a"	1′ Elevation 780′
Cardiff	164′

9. South Onondaga Quadrangle, Onondaga Co. — Case Hill Ravine 100 feet west of Case Road and up from U.S. Route 20. Type Section Case Hill Coral Bed.

Delphi Station	Not measured
Mottville	
Stereolasma Community	1.5′
limestone "b"	9′
Tropidoleptus Community	2.25′
limestone "a"	1.5′ Elevation 760′
Cardiff	138′

10. Otisco Valley Quadrangle, Onondaga Co. — Bare Mountain Ravine (Rattlesnake Gulf) on west side of Tully Valley, just north of and parallel to Otisco Road.

Delphi Station	Not measured
Mottville	
Stereolasma Community	3′
limestone "b"	8′
Tropidoleptus Community	6′
limestone "a"	1′ Elevation 680′
Cardiff	81′

11. Otisco Valley Quadrangle, Onondaga Co. — Rainbow Creek, opposite Bare Mountain Ravine on east side of Tully Valley, up from N.Y. Route 11A at its intersection with Otisco Road.

Delphi Station	Not measured
Mottville	
Stereolasma Community	3′
limestone "b"	0.5′
Tropidoleptus Community	5′
limestone "a"	1.5′ Elevation 680′
Cardiff	52′

12. Jamesville Quadrangle, Onondaga Co. — Unnamed ravine up from Apulia Road on the west side of the Butternut Creek Valley, 2.7 miles south of Jamesville.

Delphi Station	Not measured
Mottville — poor exposure	
Stereolasma Community	2′
limestone "b"	1.5′
Tropidoleptus Community	7′ only upper 2′exposed
limestone "a"	2′ Elevation 1,000′
Cardiff	188′

13. Oran Quadrangle, Onondaga Co. — Pratts Falls about 2.4 miles northwest of Pompey Center.

Delphi Station	Not measured
Mottville	
Stereolasma Community	covered
limestone "b"	8′
Tropidoleptus Community	5′
nearly barren shales	8′ mostly covered
limestone "a"	2′ Elevation 1,120′
Cardiff	7′

14. Oran-DeRuyter Quadrangles, Madison Co. — Knights Falls Ravine (unnamed on quadrangle) on the east side of Limestone Creek Valley (Pompey Hollow), 0.5 miles east of the village of Delphi Falls and 0.7 miles north of Delphi Falls Ravine.

Delphi Station	Not measured
Mottville	
Stereolasma Community	1′ Elevation 1,020′
limestone "b"	2.5′
Tropidoleptus Community	5′
nearly barren shales	42′
limestone "a"	1′ Elevation 965′

15. Oran Quadrangle, Madison Co. — Abandoned New York Central (West Shore) Railroad tunnel about 2 miles northwest of Cazenovia crossing beneath N.Y. Route 92, just north of its junction with West Lake Road.

15a East Portal — East of N.Y. 92

Mottville	
nearly barren shales	36′
limestone "a"	1.5′ Elevation 1,220′
Cardiff	1′

15b West Portal — West N.Y. 92

Mottville	
nearly barren shales	6′
limestone "a"	1′ Elevation 1,220′
Cardiff	40′
Spinocyrtia Zone (4′ thick)	19′ below limestone "a"

16. Cazenovia Quadrangle, Madison Co. — Small ravine on west side of the Chittenango Valley about 3 miles north of Cazenovia, 0.2 miles southwest of the 90 degree bend in Rathbun Road adjacent to Chittenango Falls State Park, and down from Ridge Road.

Mottville	
nearly barren shales	21′
limestone "a"	1′ Elevation 1,340′
Cardiff	190′
Chittenango	130′

17. Munnsville Quadrangle, Madison Co. — Unnamed ravine up from Whites Corners Road 0.3 miles south of Whites Corners and 0.8 miles north of its intersection with U.S. Route 20 at Morrisville Station.

Delphi Station	35′
Cole Hill Horizon	10′ Elevation 1,455′
Mottville	
sandy Mottville	7′ Elevation 1,430′
nearly barren shales	43′ (lower 18′ covered)
limestone "a"?	2′ Elevation 1,380′

18. Munnsville Quadrangle, Madison Co. — Roadcut on Crow Hill Road about 0.2 miles north of its intersection with U.S. Route 20, about 0.4 miles east of Pine Woods.

Mottville	
nearly barren shales	20′
limestone "a"	1′ Elevation 1,380′
Pecksport	160′ only upper 20′ exposed
Solsville	Elevation 1,220′
exposed on south side of U.S. 20 opposite intersection with Crow Hill Road	

19. Munnsville Quadrangle, Madison Co. — Unnamed ravine (Livermore Gully?) up from N.Y. Route 12B, 1 mile northeast of its junction with N.Y. Route 46 at Pecksport. Ravine is on the southeast side of the Chenango Valley just opposite dirt road near south edge of quadrangle.

Delphi Station	Not measured
Cole Hill Horizon	10′ Elevation 1,325′
sandy Mottville	10′ Elevation 1,350′
Pecksport	135′

20. Hamilton Quadrangle, Madison Co. — Unnamed ravine up from N.Y. Route 12B, 1.2 miles southeast of its junction with N.Y. Route 46 at Pecksport.

sandy Mottville	10′ Elevation 1,220′
Pecksport	19′

21. Hamilton Quadrangle, Madison Co. — East tributary of Payne Brook just north and parallel to Payne Avenue on eastern side of Hamilton, just east of the country club.

sandy Mottville	25′ Elevation 1,180′
Pecksport	22′

22. Munnsville Quadrangle, Madison Co. — Johnnycake Hill Ravine on the east side of Johnnycake Hill, 0.9 miles southwest of Madison.

Delphi Station	Not measured
Cole Hill Horizon	10′ Elevation 1,440′
sandy Mottville	10′ Elevation 1,420′
Pecksport	62′

23. Oriskany Falls Quadrangle, Madison Co. — Unnamed ravine down from Stone Road, 0.15 miles southwest of its intersection with Landers Road, 2.8 miles south of Oriskany Falls.

sandy Mottville	10′ Elevation 1,580′

24. Hubbardsville Quadrangle, Madison Co. — Unnamed tributary up from Tinker Hollow Road flowing northeast and joining Tinker Hollow just east of the Madison-Oneida Co. line and just east of the northerly bend in Tinker Hollow Road, about 2.85 miles northwest of North Brookfield.

Delphi Station	25′
Cole Hill Horizon	5′ Elevation 1,485′
sandy Mottville	20′ Elevation (base) 1,440′
Pecksport	80′

25. Hubbardsville Quadrangle, Madison Co. — Road cut on Cole Hill Road, 0.4 miles south of its intersection with Swamp Road, west side of Sangerfield Valley and about 2 miles west of North Brookfield.

Delphi Station	40′
Cole Hill Horizon	7′ Elevation 1,390′
sandy Mottville	6′ Elevation top 1,360′

26. Hubbardsville Quadrangle, Madison Co. — Unnamed ravine up from Fitch Road on east side of Sangerfield Valley south of Ritz Hill and north of York Hill flowing west from York Pond, 1.3 miles northwest of West Brookfield.

Delphi Station	Not measured
Cole Hill Horizon	Elevation top 1,280′
sandy Mottville (mostly covered)	3′ Elevation 1,255′

27. Cassville Quadrangle, Oneida Co. — Unnamed ravine parallel to and just south of Tubbs Road about 0.7 miles southeast of Five Corners and 0.7 miles northwest of Baily Lake.

Delphi Station	20′
Cole Hill Horizon	Elevation 1,600′
sandy Mottville	13′ Elevation 1,580′

28. Brookfield Quadrangle, Madison Co. — Road cut and quarry on the west side of Beaver Creek Valley just west of Beaver Creek Road about 2 miles north of Brookfield.

Delphi Station	50′
Cole Hill Horizon	10′ Elevation 1,535′
sandy Mottville	Top only, Elevation 1,500′

29. Brookfield Quadrangle, Madison Co. — Button Falls up and down from Button Falls Road on west side of Unadilla Valley about 1.2 miles southwest of Leonardsville.

Delphi Station	7′
sandy Mottville	10′ Elevation 1,300′
Pecksport	90′
Solsville	8′ Elevation 1,200′

30. Unadilla Forks Quadrangle, Otsego Co. — Unnamed ravine and quarry parallel to and just west of County Road 21, 0.3 miles north of Plainfield Center and south of West Winfield.

sandy Mottville	12′ Elevation 1,670′
Pecksport	70′
Solsville	10′ Elevation top 1,600′

31. Schuyler Lake Quadrangle, Otsego Co. — Roadcut on west side and in drainage ditch along County Road 24, 2.7 miles southwest of Richfield.

sandy Mottville	6′ Elevation 1,820′
Pecksport (Panther Mountain)	65′
Solsville	65′ Elevation top 1,750′

32. Schuyler Lake Quadrangle, Otsego Co. — Exposure in drainage ditch and quarry on east side of Otsego County Road 24 south of Cobblestone School at intersection of Otsego County Road 23.

Panther Mountain	140′
Delphi Station?	60′
Cole Hill Horizon?	Elevation 1,780′
Mottville?	20′ Elevation top 1,740′
Pecksport?	60′

33. Schuyler Lake Quadrangle, Otsego Co. — Lidell Creek at base of north slope of Tunnicliff Hill, just south and parallel to Otsego County Road 16 about 1.5 miles southwest of its junction with N.Y. Route 28.

Panther Mountain	180′
Cystiphylloides Coral Bed	Elevation 1,640′
base of exposure	Elevation 1,520′

Exposure is probably well above Mottville but is interesting for the coral bed containing enormous numbers of *Cystiphylloides americanum*.

34. Richfield Springs Quadrangle, Otsego Co. — Exposure in drainage ditch and small roadcut at junction of unnamed roads 0.4 miles north of Rum Hill.

Panther Mountain	60′
Mottville?	Elevation 1,880′
Paraspirifer horizon	Elevation 1,860′
Solsville	Elevation top 1,820′

35. Richfield Springs Quadrangle, Otsego Co. — Roadcut on east side of road about 1 mile south of Pierstown.

Panther Mountain	60′
Mottville?	Elevation 1,600′
Solsville	Elevation top 1,540′

APPENDIX II

MOTTVILLE FAUNAL DATA

Mottville faunal content, diversity and relative abundance arranged by feeding type. See Appendix I and Figure 1 for detailed locality information. See text, page 7 for explanation of relative abundance codes.

Limestone "a"

FEEDING TYPE AND CLASSIFICATION		LOCALITY NUMBER / TAXON	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18
INFAUNAL DEPOSIT	Anne lids	*Zoophycos*	3	3	3	3	3	4	4	4	4	4	3		2	2	3	3		
	Bivalves	*Nuculites*							1										1	
		Nuculoidea			2								1		1	2			2	
		Paleoneilo	2	2	1			1		1		2								2
RECLINED FILTER	Brachiopods	*Ambocoelia*	3	4	6	4	3	2	2	1	2	4			1		3	5		
		Chonetids	3	4	6	3	2	5	5	2		4	2		3	5	2	5	2	
		Douvillina																		
		Longispina				2														
		Mucrospirifer mucronatus				1		1	3	1		2			2	2	4		2	
		Pholidostrophia nacrea						1			1	3	1	2	1		1			
		Productella																		
		Protoleptostrophia								1		1	4	3	2		1			
		Pustulatia pustulosa							1											
		Schuchertella				1														
		Spinatrypa																		
		Tropidoleptus								1		4							1	
EPIFAUNAL, EPIBYSSATE, AND SHALLOW ENDOBYSSATE FILTER	Brachiopods	*Allanella tullius*						1	1			1								
		Athyris cora																		
		Athyris spiriferoides																		
		"*Camarotoechia*"																		
		Elytha fimbriata																		
		Leiorhynchus	3	2			1	3	1								2			
		Mediospirifer audaculus						4	4	4	3	4	3	2		3	4	3	3	3
		Meristella																		
		Nucleospira						2	2	2	1							4		
		Paraspirifer																		
		Rhipidomella						1				3								
		Spinocyrtia																		
		Spirifer sculptilis																		
	Bivalves	*Actinopteria*			1								2							
		Glyptodesma											1							
		Leiopteria										1								
		Mytilarca											1							
		Pectenids										2								
		Pseudaviculopecten						1												
		Pterinopecten										1	3							
		Ptychopteria (*Cornellites*)											2							
	Bryozoans	Bryozoans	1 ?																	
		Fenestella																		
INFAUNAL FILTER	Bivalves	*Cypricardella*																		
		Goniophora								1	1	1								
		Gosseletia											3							
		Grammysia									1	1	1							
		Nacrodon																		
		Modiomorpha concentrica										1	1						1	
		Modiomorpha mytiloides							1	1		1	2							
		Nyassa																		
		Paracyclas							1											
		Pholadella						1												
MICROCARNIVORES	Tabulates	*Aulocystis*				2														
		Favosites alpenensis																		
		F. arbuscula																		
		F. argus								1	1									
		F. hamiltoniae								1	1									
	Rugose	*Amplexiphyllum*																		
		Cystiphylloides																		
		Eridophyllum																		
		Heliophyllum halli							2	2	3	2	1							
		Heterophrentis									1									
		Stereolasma						1	1											
EPIFAUNAL MOBILE COLLECTORS	Gastropods	*Bellerophon*																		
		Bembexia			2							1	. 2		3	2	2			
		Hourlonia itys			1	1			2			1								
		Paleozygopleura (*Loxonema*)				1									2					
		Platyceras				2							1							
	Trilobites	*Dipleura dekayi*	1	2		2	1	1	2			2	3	2	2		3	3		1
		Greenops boothi	1		1	1		1	1						1					
		Phacops rana	2	2	2	4	4	2	3			3		2	2	2	3			
		Proetus																		
NEKTONIC CARNIVORES	Cephalopods	*Orthoceras*	2	3	1		2		2						1					
		Tornoceras			1			1	1											
EPIPLANKTONIC FILTER	Bivalves	*Panenka*	1																	
		Pterochaenia		1				3	1											

Tropidoleptus Community

FEEDING TYPE AND CLASSIFICATION		LOCALITY NUMBER TAXON	1	2	3	4	5	6	7	8	9	10	11	12	13	14
INFAUNAL DEPOSIT	Annelids	*Zoophycos*										4	2			
	Bivalves	*Nuculites*									1		2			2
		Nuculoidea									1	2				2
		Paleoneilo							1	1	1	3	3			3
RECLINED FILTER	Brachiopods	*Ambocoelia*							1			6	1	5		1
		Chonetids							5	3	4	3	4	5	4	4
		Douvillina														
		Longispina										2				
		Mucrospirifer mucronatus							1		2	2	2			2
		Pholidostrophia nacrea										3	1			
		Productella														1
		Protoleptostrophia									1	1	2			
		Pustulatia pustulosa							1				2			
		Schuchertella							1							
		Spinatrypa														
		Tropidoleptus							2	1	4	5	4	4	6	6
EPIFAUNAL, EPIBYSSATE, AND SHALLOW ENDOBYSSATE FILTER	Brachiopods	*Allanella tullius*								1	1	1				
		Athyris cora														
		Athyris spiriferoides												1		
		"*Camarotoechia*"									1					
		Elytha fimbriata								2		1				
		Leiorhynchus							1	1	1		1			
		Mediospirifer audaculus								1	4	4	3		1	3
		Meristella														
		Nucleospira							1		3	3	1			
		Paraspirifer														
		Rhipidomella							1		1	1	1			
		Spinocyrtia														
		Spirifer sculptilis										2				
	Bivalves	*Actinopteria*									1		1			2
		Glyptodesma														
		Leiopteria											2	4		
		Mytilarca														
		Pectenids											1		2	3
		Pseudaviculopecten									2					
		Pterinopecten														2
		Ptychopteria (*Cornellites*)										1	3			3
	Bryozoans	Bryozoans										2				
		Fenestella														1
INFAUNAL FILTER	Bivalves	*Cypricardella*										1		1		
		Goniophora									1	1	1			
		Gosseletia														
		Grammysia														3
		Macrodon								1		3	1			3
		Modiomorpha concentrica										1	1			2
		Modiomorpha mytiloides								1	1	2				
		Nyassa														1
		Paracyclas									1	1				
		Pholadella							1			2				1
MICROCARNIVORES	Tabulates	*Aulocystis*														
		Favosites alpenensis														
		F. arbuscula														
		F. argus														
		F. hamiltoniae														
	Rugose	*Amplexiphyllum*														
		Cystiphylloides														
		Eridophyllum														
		Heliophyllum halli							1							
		Heterophrentis														
		Stereolasma									1					
EPIFAUNAL MOBILE COLLECTORS	Gastropods	*Bellerophon*										1				
		Bembexia							2	1	2	2	1			1
		Mourlonia itys								1	1		1			
		Paleozygopleura (*Loxonema*)							1			2	1			2
		Platyceras														
	Trilobites	*Dipleura dekayi*											2		1	
		Greenops boothi								1	1	1				2
		Phacops rana							1		1	1				
		Proetus														
NEKTONIC CARNIVORES	Cephalopods	*Orthoceras*							1			1	1			1
		Tornoceras								1	1		1			
EPIPLANKTONIC FILTER	Bivalves	*Panenka*														
		Pterochaenia														

Zoophycos Swirl-Rock and *Heliophyllun-Favosites* Community (Limestone "b")

FEEDING TYPE AND CLASSIFICATION		LOCALITY NUMBER	1	2	3	4	5	6	7	8	9	10	11	12	13	14
		TAXON														
INFAUNAL DEPOSIT	Annelids	*Zoophycos*	6	5	5	5	5	5	4	4	4	4	3	3	3	3
	Bivalves	*Nuculites*														
		Nuculoidea														
		Paleoneilo												1		1
RECLINED FILTER	Brachiopods	*Ambocoelia*														1
		Chonetids									2	2			4	3
		Douvillina														1
		Longispina														
		Mucrospirifer mucronatus										2	1			2
		Pholidostrophia nacrea									1	2				2
		Productella														
		Protoleptostrophia												3	1	
		Pustulatia pustulosa														
		Schuchertella														
		Spinatrypa														
		Tropidoleptus												3		3
EPIFAUNAL, EPIBYSSATE, AND SHALLOW ENDOBYSSATE FILTER	Brachiopods	*Allanella tullius*														
		Athyris cora														1
		Athyris spiriferoides											1			
		"*Camarotoechia*"														
		Elytha fimbriata														1
		Leiorhynchus													1	1
		Mediospirifer audaculus														
		Meristella														
		Nucleospira									1	2	1			
		Paraspirifer														
		Rhipidomella									2	3	1	3		2
		Spinocyrtia														
		Spirifer sculptilis														
	Bivalves	*Actinopteria*											1	2		
		Glyptodesma														1
		Leiopteria														
		Mytilarca														1
		Pectenids												1	1	2
		Pseudaviculopecten													2	
		Pterinopecten														2
		Ptychopteria (Cornellites)													3	3
	Bryozoans	Bryozoans								1	1					
		Fenestella										2			4	4
INFAUNAL FILTER	Bivalves	*Cypricardella*														
		Goniophora														
		Gosseletia														
		Grammysia														
		Macrodon														
		Modiomorpha concentrica														2
		Modiomorpha mytiloides														
		Nyassa														
		Paracyclas														2
		Pholadella														
MICROCARNIVORES	Tabulates	*Aulocystis*														
		Favosites alpenensis									2	2				
		F. arbuscula								1	2	2	2			
		F. argus								2	3	2				
		F. hamiltoniae									2	2				
	Rugose	*Amplexiphyllum*										2				
		Cystiphylloides									1	2				
		Eridophyllum									1					
		Heliophyllum halli								3	5	3	2		1	1
		Heterophrentis								1	2	2		3	3	2
		Stereolasma								2	1					
EPIFAUNAL MOBILE COLLECTORS	Gastropods	*Bellerophon*														
		Bembexia												1	3	
		Mourlonia itys								1		1			2	1
		Paleozygopleura (Loxonema)										2			1	2
		Platyceras														
	Trilobites	*Dipleura dekayi*							1				1		2	
		Greenops boothi														
		Phacops rana								1		1		1	2	
		Proetus														
NEKTONIC CARNIVORES	Cephalopods	*Orthoceras*		2		2			2						2	2
		Tornoceras														
EPIPLANKTONIC FILTER	Bivalves	*Panenka*														
		Pterochaenia														

Stereolasma Community

FEEDING TYPE AND CLASSIFICATION		LOCALITY NUMBER TAXON	1	2	3	4	5	6	7	8	9	10	11	12	13	14
INFAUNAL DEPOSIT	Annelids	*Zoophycos*	6	5	5		5									
	Bivalves	*Nuculites*						1								2
		Nuculoidea														1
		Paleoneilo					1					1	1			3
RECLINED FILTER	Brachiopods	*Ambocoelia*					2	4	3	1	2	3	2	4		2
		Chonetids					2		1	3	2	2	1	4		4
		Douvillina					1	1	1							
		Longispina														
		Mucrospirifer mucronatus					1									1
		Pholidostrophia nacrea						2			2	2	2			2
		Productella						1	1		1					1
		Protoleptostrophia									1	1	1	3		2
		Pustulatia pustulosa														1
		Schuchertella														
		Spinatrypa								1						
		Tropidoleptus														
EPIFAUNAL, EPIBYSSATE, AND SHALLOW ENDOBYSSATE FILTER	Brachiopods	*Allanella tullius*														
		Athyris cora														
		Athyris spiriferoides														
		"*Camarotoechia*"														
		Elytha fimbriata														
		Leiorhynchus							1			1				
		Mediospirifer audaculus						1								2
		Meristella						1								
		Nucleospira						1	2	1	1	2	1			
		Paraspirifer														
		Rhipidomella					1	4	3	3	3	4	2	1		2
		Spinocyrtia														
		Spirifer sculptilis														
	Bivalves	*Actinopteria*														1
		Glyptodesma														
		Leiopteria														
		Mytilarca														
		Pectenids														3
		Pseudaviculopecten														
		Pterinopecten														
		Ptychopteria (*Cornellites*)														
	Bryozoans	Bryozoans					1				2					2
		Fenestella								1		2				5
INFAUNAL FILTER	Bivalves	*Cypricardella*					1									2
		Goniophora														
		Gosseletia														
		Grammysia														
		Macrodon														
		Modiomorpha concentrica														
		Modiomorpha mytiloides									1					
		Nyassa														2
		Paracyclas						1	1							2
		Pholadella														
MICROCARNIVORES	Tabulates	*Aulocystis*					1	1	1	1		1				
		Favosites alpenensis														
		F. arbuscula														
		F. argus														
		F. hamiltoniae														
	Rugose	*Amplexiphyllum*					1	5	5	4	1	1				
		Cystiphylloides														
		Eridophyllum														
		Heliophyllum halli												1		
		Heterophrentis														
		Stereolasma					2	6	6	5	4	3	3	2		3
EPIFAUNAL MOBILE COLLECTORS	Gastropods	*Bellerophon*														
		Bembexia					1	1	2	2	1	1	1			
		Mourlonia itys					1	1			1	2	1	1		
		Paleozygopleura (*Loxonema*)					1			1	2	2	1			
		Platyceras						4		1	1					
	Trilobites	*Dipleura dekayi*														
		Greenops boothi					1	1	1		1		1			2
		Phacops rana					5	1	4	3	1	3	1	1		3
		Proetus								1						
NEKTONIC CARNIVORES	Cephalopods	*Orthoceras*		2	2		4	2		1	1	1	1			
		Tornoceras					1	1	1	1						
EPIPLANKTONIC FILTER	Bivalves	*Panenka*														
		Pterochaenia														

"*Camarotoechia*"-*Allanella* Community

FEEDING TYPE AND CLASSIFICATION		LOCALITY NUMBER	17	18	19	20	21	22	23	24	25	26	27	28	29	30	31	32	34	35
		TAXON																		
INFAUNAL DEPOSIT	Annelids	*Zoophycos*	2		2	3	2	3	2	3		2	3		3	4				
	Bivalves	*Nuculites*					1													
		Nuculoidea					1						1			4				
		Paleoneilo	3				1									4				
RECLINED FILTER	Brachiopods	*Ambocoelia*	3													6				
		Chonetids	4			2							2					1		
		Douvillina																		
		Longispina																		
		Mucrospirifer mucronatus	3		4	3	3	3	3	3		2	3		3	3	3	3		
		Pholidostrophia nacrea																		
		Productella	3			2		2												
		Protoleptostrophia			2			2				3				1				
		Pustulatia pustulosa															2			
		Schuchertella																		
		Spinatrypa						2												
		Tropidoleptus	3					3					1		3			1		2
EPIFAUNAL, EPIBYSSATE, AND SHALLOW ENDOBYSSATE FILTER	Brachiopods	*Allanella tullius*	4		5	5	5	5	5	3			4			5	6	5	3	4
		Athyris cora																		
		Athyris spiriferoides																		
		"*Camarotoechia*"	6		6	5	6	4	6	5		4	6		5	6	6	6	4	5
		Elytha fimbriata																		
		Leiorhynchus											2			2				
		Mediospirifer audaculus	3				3		4						2					
		Meristella																		
		Nucleospira																		
		Paraspirifer	2 ?												2	2			3	2
		Rhipidomella						3									1			
		Spinocyrtia					2	1	1	1										
		Spirifer sculptilis				1														
	Bivalves	*Actinopteria*	1										1							
		Glyptodesma																		
		Leiopteria																		
		Mytilarca				2									2	2	2	3		
		Pectenids	3					1	2	3		3				3	2			
		Pseudaviculopecten					1													
		Pterinopecten																		
		Ptychopteria (*Cornellites*)	4		1	3	2	2	1	1					2	1	2			
	Bryozoans	Bryozoans																		
		Fenestella					1													
INFAUNAL FILTER	Bivalves	*Cypricardella*													5	4		3		
		Goniophora	1					1								1				
		Gosseletia																		
		Grammysia	2					1					2			1	1			
		Macrodon	2					1												
		Modiomorpha concentrica	2			1	1	1		1			1				1			1
		Modiomorpha mytiloides																		
		Nyassa																		
		Paracyclas											1						3	
		Pholadella	2																	
MICROCARNIVORES	Tabulates	*Aulocystis*																		
		Favosites alpenensis																		
		F. arbuscula																		
		F. argus																		
		F. hamiltoniae	3					2								1	2			
	Rugose	*Amplexiphyllum*																		
		Cystiphylloides																		
		Eridophyllum																		
		Heliophyllum halli				3	3	3												
		Heterophrentis				3											3		1	
		Stereolasma																		
EPIFAUNAL MOBILE COLLECTORS	Gastropods	*Bellerophon*			1															
		Bembexia	2																	
		Mourlonia itys																		
		Paleozygopleura (*Loxonema*)													2					
		Platyceras																		
	Trilobites	*Dipleura dekayi*	2		2	4	1	2	2	1			2		2	2	2	1	2	
		Greenops boothi	2												2					
		Phacops rana	2																	
		Proetus																		
NEKTONIC CARNIVORES	Cephalopods	*Orthoceras*			1	1														
		Tornoceras																		
EPIPLANKTONIC FILTER	Bivalves	*Panenka*																		
		Pterochaenia																		

Faunal and Lithologic Cyclicity in the Centerfield Member (Middle Devonian: Hamilton Group) of Western New York: A Reinterpretation of Depositional History

Michael Savarese,[1] Lee M. Gray,[2] and Carlton E. Brett
Department of Geological Sciences
The University of Rochester

ABSTRACT

The Centerfield Member (Middle Devonian, Givetian) in western New York constitutes a nearly symmetrical cycle of marine litho- and biofacies, representing gradational changes in depositional environments. This member comprises a central limestone which grades both above and below into fossiliferous, gray calcareous claystones and ultimately into dark-gray, sparsely fossiliferous shales. In the Genesee Valley (Livingston County), the entire Centerfield sequence is well exposed, allowing detailed, stratigraphically controlled faunal sampling. Both presence-absence and absolute-abundance fossil data from bulk-samples were analyzed using weighted-pair group-averaging cluster-analysis. The distribution of biofacies and patterns of faunal diversity and relative abundance all support the existence of cyclicity. The Centerfield Member apparently records a regressive-transgressive cycle wherein the middle limestone was deposited in a shallow-water setting and the surrounding dark shales were deposited in deeper, offshore environments. Taphonomic, sedimentologic, and stratigraphic evidence further corroborate this interpretation. The existence of other limestone-centered cycles within the Hamilton Group (e.g., Stafford-Mottville, Tichenor-Portland Point) with identical faunal patterns suggests that several regressive-transgressive cycles, similar to that of the Centerfield, occurred during Givetian time.

INTRODUCTION

The recognition and environmental reconstruction of marine cyclothems has been based principally on sedimentologic and stratigraphic evidence (Heckel, 1977; Kauffman, 1977). Little emphasis has been placed on the effects that these transgressive-regressive fluctuations have had on the associated fauna (but see von Bitter, 1972; Merrill and von Bitter, 1976). The Hamilton Group (Middle Devonian) of western New York is relatively homogeneous in lithology (mainly mudstones) and generally lacks sedimentologic features useable for environmental reconstruction. Recently, however, a number of depositional cycles have been recognized in rocks of the Hamilton Group based largely on faunal distribution (Brett and others, 1983). Faunal associations have been described previously for the Hamilton Group of New York in numerous studies (Fernow, 1961; Beerbower, et al., 1969; Grasso, 1970, 1973, this volume; Brett, 1974a, 1974b; Koch, 1978; McCollum, 1980; Miller, 1982, this volume), but few attempts have been made to integrate shifts in faunal associations with transgressive-regressive fluctuations.

The Hamilton Group spans latest Eifelian to middle Givetian time and its strata probably represent 6-7 million years of deposition (Harland and others, 1982). In western New York State it is positioned between the lower Middle Devonian Onondaga Limestone and the Upper Devonian Genesee Group. The Hamilton of central and western New York is divisible into two broad facies: (1) the "Marcellus facies," comprised of gray to black sparsely fossiliferous shales, which dominates the lower Marcellus and Skaneateles Formations, and (2) the "Moscow facies," composed of fossiliferous gray mudstones and thin calcareous units which forms the bulk of the upper Ludlowville and Moscow Formations (Cooper, 1957; Rickard, 1975). Both of these facies grade eastward into silty "Hamilton facies" and finally into red beds ("Catskill facies").

The four formations of the Hamilton Group — the Marcellus, Skaneateles, Ludlowville, and Moscow — were established by means of widespread carbonate-rich beds. One of these intervals, the Centerfield Member, that marks the base of the Ludlowville Formation has been considered a "prime key bed" of the Hamilton (Cooper, 1930; Rickard, 1975, 1981). The present study deals with the Centerfield Member in a restricted geographic region of western New York State.

The Hamilton Group, although described early in the history of New York State geology, has been little studied in terms of its depositional history. Interest in the Hamilton Group was eclipsed by studies of the seemingly more complex history of the Upper Devonian rocks overlying the Hamilton. Waltherian

---

[1] Present address: Dep't of Geology, University of California at Davis

[2] Present address: Dep't of Geology, Mt. Union College, Alliance, Ohio

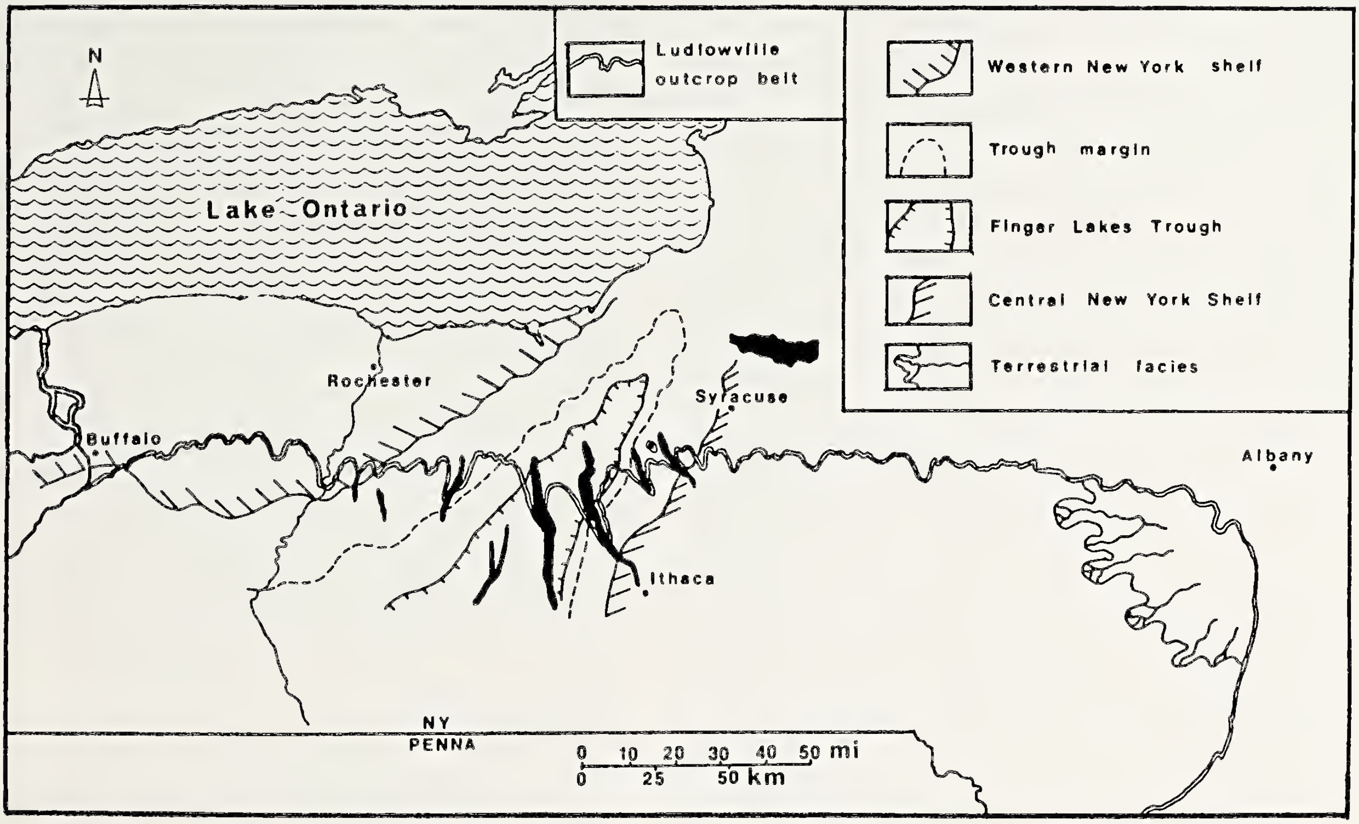


Figure 1: A general paleogeographic reconstruction map of the northern Appalachian Basin during Givetian time. The Ludlowville Formation outcrop belt is included.

lithofacies relationships were easily recognized in outcrops with obvious eastward shallowing trends associated with the Catskill deltaic complex. Recently, however, the interpretation of depositional history of the Hamilton beds has been reevaluated. Subtle facies shifts within members are now apparent in western and central New York as well as the more obvious eastward coarsening trend.

A recently proposed paleogeographic reconstruction of the northern Appalachian Basin (Fig. 1) helps to explain Hamilton facies relationships (Brett, Baird, and Kloc, 1983). The orientation of the basin axis and slope, combined with the fortuitous parallelism of the Hamilton outcrop belt (Fig. 1) and the ancient facies tracts in western New York, result in an apparent layer-cake stratigraphic sequence. A number of recent studies have provided detailed facies evidence for an east-west orientation of facies belts in several rock units of Silurian and Devonian age in western New York (Belak, 1980; Kissling and Moshier, 1981; Brett, 1983). A broad shelf sea existed across most of central and western New York (Brett, Baird and Kloc, 1983); however, a simple east-west deltaic profile, as often interpreted for Upper Devonian sequences, did not exist during Middle Devonian time. Differential subsidence in the western Finger Lakes region allowed thick, anoxic deposits to accumulate along the basin's northeast-southwest trending axis (Baird, 1981; Baird and Brett, 1981). The NE-SW oriented axis of subsidence extended southward into western Pennsylvania, and West Virginia (Dennison and Hasson, 1976; Dennison and Head, 1975; Baird and Brett, 1981). Depositional history in the northern Appalachian Basin was further complicated by a shift in the position of its axis through Hamilton time. Instead of remaining static, the deepest portion of the basin slowly migrated westward (Ettensohn and Barron, 1981, 1982; Brett, Baird, and Kloc, 1983; Brett and Baird, 1982).

East of the axial region, the basin sloped gradually toward shallower depths (Fig. 1). This eastern shelf with its slope running parallel to the northeast-southwest trending axis, was closer to an early Acadian clastic source and therefore accumulated coarser-grained sediments (Baird and Brett, 1981). A stable shelf region also existed to the west (sometimes called the "Buffalo Arch or Axis"; Brett and Baird, 1982). Low-lying, shallow subtidal areas to the north and west accumulated limestones and calcareous mudstones. Following this model, the coincidence between the strike of the northward shoaling facies belts and that of the present Hamilton outcrop belt explains the near lack of facies transitions for Hamilton strata in western New York. If the orientation of the western shelf had been perpendicular to the strike of the outcrop belt instead of parallel to it, a continuous east-to-west litho- and biofacies shift within Hamilton units would exist. In fact, subtle facies shifts can be recognized in western New York when the outcrop belt arcs northward or southward, and hence cuts obliquely across the inferred paleoslope. These changes appear as minor shifts in bio- and/or lithofacies. Abrupt facies shifts can be observed along the outcrop belt eastward from the central Finger Lakes, and, thus, upslope from the inferred basin center onto the eastern clastic shelf.

Vertical sequences of lithofacies observed in the Hamilton Group evidently reflect fluctuations in depositional environment. Certain authors have invoked simple oscillations of variables such as oxygen level and sedimentation rate without change in relative depth as the primary cause of this facies change through time (Cooper, 1957; McCollum, 1980). However, detailed regional correlation of Hamilton units and apparent lateral facies gradations east of the Finger Lakes (Baird, 1979; Baird and Brett, 1981, 1983a; Brett and Baird, 1985) have led to the proposal of an alternative, Waltherian explanation: vertical changes in single sections of western New York reflect lateral migration of elongate facies belts in a northward or southward direction as a result of transgressive-regressive cycles (Brett and Baird, 1982; Baird and Brett, 1983a). Such facies migration for the Hamilton Group of western New York provides a plausible explanation for the gross pattern of lithofacies distribution observed in the outcrop belt. This model, however, requires further documentation and testing at both regional and local scales.

If the vertical sequence of changing lithologies resulted from lateral migration of facies belts roughly perpendicular to the outcrop belt, two effects should be observable at the scale of single outcrops. First, litho- and biofacies should show gradual transitions through vertical sequences. At the level of a single outcrop such an effect would actually be consistent with either a transgressive-regressive model or a vertical stacking ("layer-cake") model for facies change. A general lack of gradational sequences, however, would potentially falsify the transgression-regression model. This possibility is particularly important to test in the aforementioned key carbonate beds, such as the Centerfield. Such units have previously been interpreted as the result of dyscyclic interruptions of sedimentation which cross-cut facies patterns, rather than as normal parts of facies spectra (Cooper, 1957). This would imply disconformable rather than gradational contacts for the limestones.

A second prediction of the facies migration hypothesis is that litho- and biofacies should follow one another in a predictable, repeating sequence. Regularly recurring sequences, without sudden, dramatic reversals would be predicted by the facies migration model. Further, transgressive and regressive episodes should give rise to cyclic patterns of recurring facies. Previously, several researchers have noted recurring faunas or fossil associations (Grabau, 1898, 1899; Cleland, 1903). Recently, apparent cyclicity has been qualitatively documented at several levels within the Hamilton Group (Brett and others, 1983). None of these studies (with the exception of Miller, this volume), however, have attempted to examine in detail the gradational sequences which link various faunal associations together. Previously, authors have dealt either at a broad scale involving hundreds of meters of sediment (Cleland, 1903; Smith, 1935) or at the scale of a single bedding plane (Flessa and Bray, 1977; Beerbower and others, 1969). Consequently, intermediate-level faunal replacement phenomena have been largely overlooked. Moreover, no previous studies (except Miller, this volume) have rigorously considered the possibility of cyclicity in patterns of fossil distribution.

The present study provides a detailed qualitative and quantitative cluster-analysis of faunal distribution through the Centerfield interval for a limited region of western New York. It is hoped that this combined approach will help to elucidate the complex faunal pattern seen in this stratigraphic interval. The objectives of this research are as follows: (1) to document the sequence of faunas through a black shale-carbonate-black shale transition; (2) to determine whether or not the fossil sequence is gradational above and below into the limestone bed; (3) to test the sequence for cyclicity of recurrent fossil associations; and (4) if such cyclicity exists, to determine its probable causes.

STRATIGRAPHY

The Centerfield Member (Ludlowville Formation) is one of several shale-limestone-shale sequences found in the Hamilton Group of western and central New York State, most of which display parallel features. The Centerfield in western New York appears to represent a relatively complete, gradational lithologic interval. It is also very fossiliferous and contains representatives of most fossil associations found in the Hamilton Group as a whole. The Centerfield Member, therefore, provides an ideal test case for lithologic and faunal cyclicity and perhaps can serve as a model for many other Hamilton carbonate sequences.

The Centerfield previously was placed at the top of the Skaneateles Formation based on the locally disconformable contact between the limestone and the overlying Ledyard Shale in the Cayuga Lake region (Smith, 1935). As presently defined, however, it marks the base of the Ludlowville Formation (Cooper and others, 1942). In the Genesee Valley region, the Centerfield is tripartite, comprised of a lower shale, a middle limestone and calcareous claystone, and an upper shale submember (Gray, 1983, 1984). The Centerfield, at its type locality on Schaeffer Creek, consists primarily of the lower shale and middle limestone submembers; the upper shale has been removed by erosion.

The Genesee River Valley in Livingston County, New York, was chosen for detailed faunal analysis of the Centerfield interval; in this area tributaries of the Genesee River have cut many sections through Hamilton Group bedrock. This study area was chosen for a number of reasons: (1) the entire Centerfield interval (4-5 m) is represented in a composite of three closely spaced outcrops; (2) a series of key beds allows for detailed correlation between localities and precise location of samples; (3) the Centerfield Member in this region represents an interval of apparently continuous deposition without major discontinuities; and (4) the Centerfield in this part of New York State is composed primarily of soft calcareous mudstone and a few limestones and, therefore, permits easy sampling. Eastward, the Centerfield becomes silty and distinctly harder and more difficult to sample.

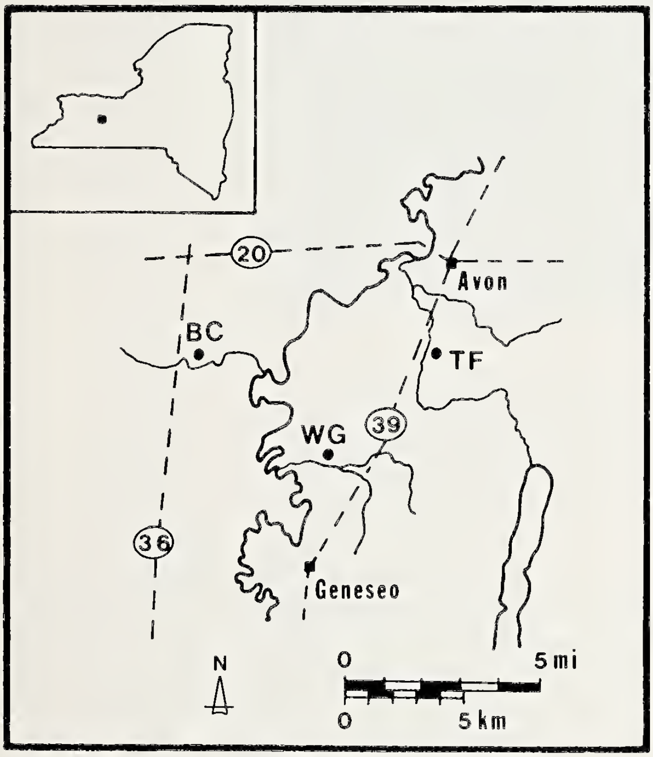


Figure 2: **Study area within the Genesee Valley, Livingston County, New York. The three localities appear on tributary streams of the Genesee River; BC, Browns Creek; WG, Wheeler Gully; TF, Triphammer Falls. Three highways are represented by dashed lines.**

All three localities are within a five-mile radius of each other and are located on the Geneseo 7 1/2' Quadrangle map, Livingston County, New York (Fig. 2).

From east to west, the three localities are:

1. Triphammer Falls (TF):Bank and waterfall exposure in Conesus Creek, approximately 0.5 km east of N.Y. Route 39, 0.3 km north of the stream's intersection with Triphammer Road, south of the town of Avon.
2. Wheeler Gully (WG):Bank and streambed exposure along Jaycox Creek, approximately 0.6 km west of N.Y. Route 39, north of the city of Geneseo.
3. Browns Creek (BC):Bank and streambed exposure along Browns Creek, approximately 1.0 km east of N.Y. Route 36, due east of the town of York (Fig. 3).

The lower half of the Centerfield Member can be seen at both Browns Creek and Triphammer Falls, whereas the upper half of the interval is exposed at Wheeler Gully. Figure 4 provides a composite stratigraphic section and lithologic description of the interval studied. Thicknesses and stratigraphic positions of the lithologies illustrated here correspond almost exactly with those measured in a drill core from the International Salt Company in the town of Retsof, approximately seven kilometers south of Browns Creek.

The lithologies present in the study area are mostly clayshales, calcareous claystones, and thin bedded argillaceous limestones. The lowest sampled beds consist of dark-gray, fissile clayshales at the top of the Levanna Member of the Skaneateles Formation. Fossils in this interval are locally abundant and tend to occur along bedding planes. The Levanna shales grade rapidly upward into moderately fossiliferous, dark-gray claystones of the "lower shale transition" (York submember of Gray, 1984) of the Centerfield Member. These, in turn, grade upwards into increasingly fossiliferous bioturbated, light-gray claystones.

Interbedded with the lower gray claystones are two thin beds of argillaceous limestone, the Salt Creek and Browns Creek beds (Gray, 1984, p. 19, 20; Figs. 3, 4). These beds are 6-10 cm thick, intensely bioturbated, and show gradational contacts with the surrounding claystones. They are quite persistent, extending laterally over approximately 100 km of outcrop belt with very little change in thickness. Fossil preservation is excellent in these beds; they have many completely articulated and uncrushed specimens of delicate fossils (trilobites, crinoids, bryozoans). These limestones may represent distal event beds produced by episodes of rapid burial by storm-suspended mud layers (see below). They are, therefore, very different in origin than the other Centerfield limestone beds.

Overlying the upper thin limestone bed (Browns Creek bed) is a 40 cm interval of extremely fossiliferous, light-gray claystone, termed the Triphammer beds by Gray (1984). Above these claystones is a 75 cm interval of argillaceous limestone with very thin shaley partings; this is the middle limestone or Schaeffer Creek submember (Gray, 1984). The rock is best described as a calcisiltite (sensu Lindholm, 1969). Included are large fossil clasts supported in a matrix of silt-sized skeletal fragments and pellets and argillaceous muds. Kramers (1970) has determined the calcium carbonate content of this middle limestone to range from 70-83%. The limestone is intensely bioturbated with no primary sedimentary structures present. Fossils generally are less abundant in the limestone than in the surrounding claystones.

Overlying the middle limestone are abundantly fossiliferous light-gray claystones (Crooked Creek coral bed) which are coral-rich and lithologically identical to the claystones below the central limestone. These claystones then grade upwards into increasingly dark-gray and more sparsely fossiliferous claystones. The "upper shale transition" (Wheeler Gully submember of Gray, 1984, p.35) is very similar both lithologically

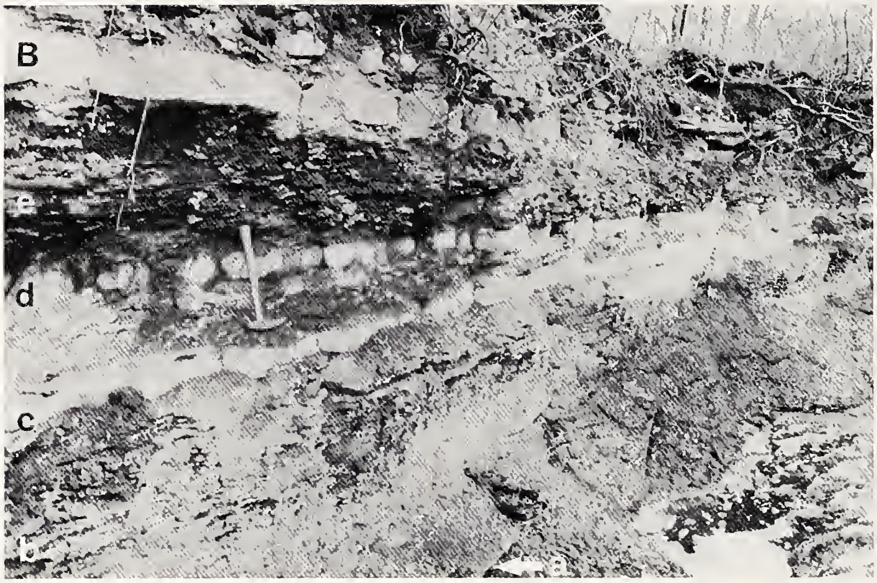


Figure 3: Outcrop of Centerfield Member on north bank of Browns Creek, York, New York (loc. 3); hammer gives scale. Letters correspond to those of Fig. 4; a) approximate position of contact between Levanna and Centerfield members; b) unnamed nodular limestone; c,d) Salt Creek and Browns Creek limestone beds; e) Triphammer interval; B) middle limestone submember of the Centerfield; upper transitional beds are missing at this locality.

and faunally to the aforementioned transitional claystones found lower in the section. The upper contact of the Centerfield Member with the overlying Ledyard Shale is lithologically gradational but can be recognized by an abrupt decrease in fossil abundance and a much darker color.

Although a lithologic cycle is recognized, the cycle of facies can be documented with greater precision by examining the fossils within the rocks. Lithologic homogeneity of most of the Centerfield necessitates both detailed qualitative and quantitative faunal analyses to interpret depositional history.

METHODS AND PROCEDURES

Microstratigraphic faunal sampling was done throughout the 5 meters of section at the three localities. Sampling sites were generally dictated by the accessibility of continuous sections, although spot checks of the same beds along outcrops indicates little lateral variation of fossil content. Bulk samples were taken at regularly spaced 10-cm intervals from the claystone and clayshale portions of the section. These intervals appear to be faunally homogeneous and finer subdivision of the section is probably unwarranted due to depth of mixing by burrowers (deepest burrows are about 10 cm). When bulk sampling was not possible, as within the limestone beds, counts were taken from blocks. In all, 34 sampling intervals were analyzed in detail (Fig. 4). The upper portion of the Centerfield was sampled at Wheeler Gully (samples 21-34) and the lower at Triphammer Falls (samples 1-18). The lower limestones along with the middle limestone (samples 11, 13, and 19) were sampled at Browns Creek (Fig. 3). The lower portion of the section also was bulk sampled at Browns Creek as a check on faunal consistency across the Genesee Valley. Furthermore, qualitative sampling of Centerfield in all nearby localities indicates a high degree of facies consistency within 20 km east and west of the Genesee Valley.

Bulk sample blocks were disaggregated in the laboratory, all fossils were identified to genus and counted, and taphonomic observations were made. Counts were adjusted uniformly for fragmented and single valve occurrences by multiplying through by a ratio (four fragments or two valves equal to one occurrence; when pedicle-versus brachial-valve distinctions were made, the higher of the two was used). The bulk-sample counts were then standardized by sample weight. Percentage and relative abundances were calculated from absolute abundances for the most common species in each sampled interval ($a<1\%$:present; $1<a<10\%$: common; $10<a<50\%$: abundant; $a>50\%$: very abundant). Shannon-Weaver dominance diversity and equitability indices were calculated for each sample (Shannon and Weaver, 1963; Lloyd and Ghelardi, 1964; Pielou, 1966; Beerbower and Jordan, 1969).

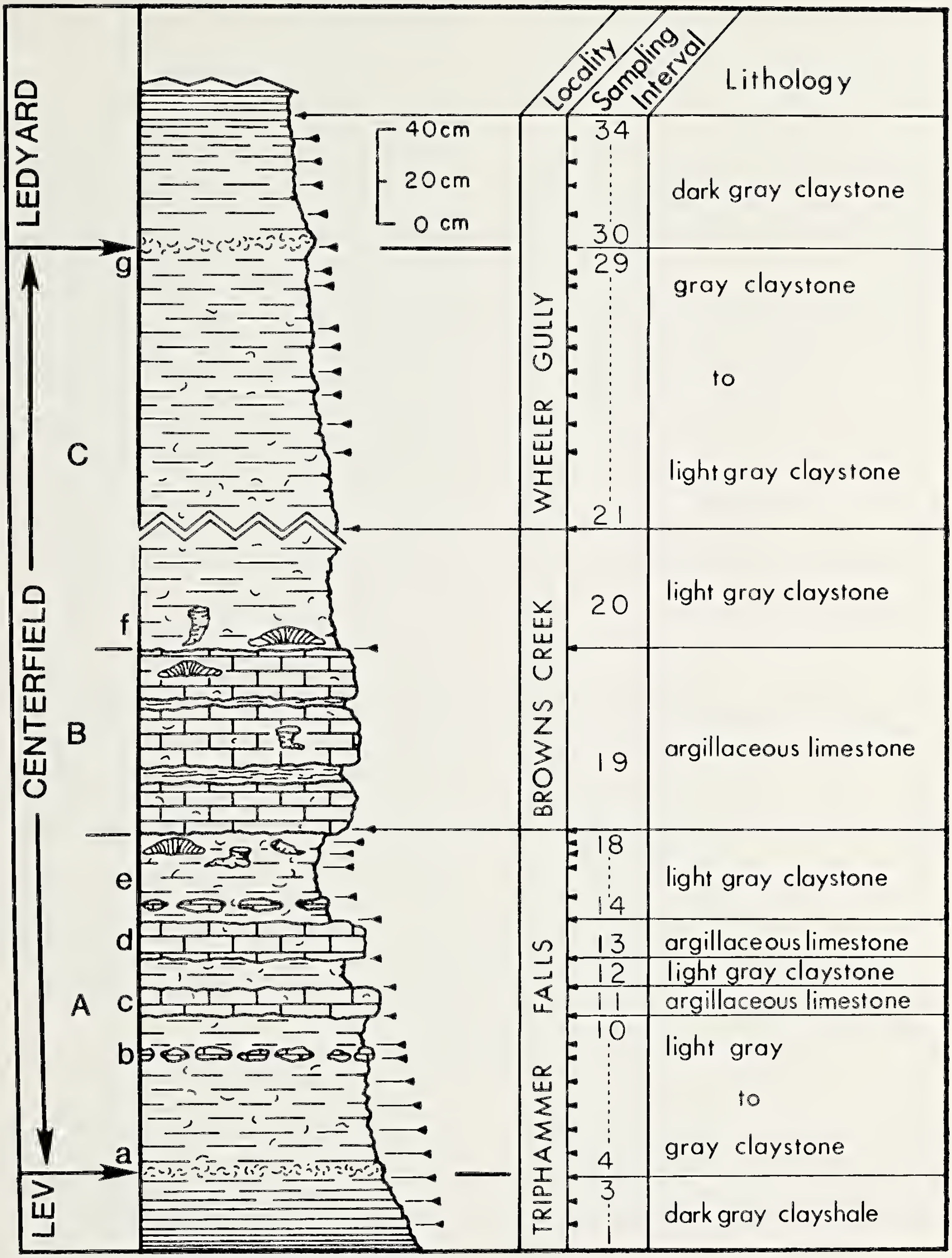


Figure 4: Centerfield composite section for the Genesee Valley with the associated lithologies found throughout the stratigraphic interval. The 34 sampling horizons are marked by the arrows alongside the section. Stratigraphic contacts and beds follow the designations of Gray (1984); LEV = Levanna Shale Member. Beds and subdivisions of Centerfield Member, designated by letters, include: A, lower transitional shales = York submember; B, middle limestone = Schaeffer Creek submember; C, upper transitional shales = Wheeler Gully submember; a, basal *Echinocoelia* – rich bed; b, unnamed nodular limestone; C, Salt Creek limestone bed; d, Browns Creek limestone bed; e, Triphammer coral bed; f, Crooked Creek coral bed; g, uppermost shell bed (probably equivalent to Kashong Glen bed).

Counts from the limestone beds were made in the field along weathered lateral surfaces of limestone beds and by breaking down blocks with sledgehammers. These data could not be treated consistently with the bulk-sample data and, therefore, diversity and equitability indices could not be calculated. However, relative abundances were determined for these samples.

Biofacies were recognized partly on the basis of cluster analysis. Faunal associations were also defined by semi-quantitative observations made after fossil counts for the 34 samples were tabulated and these were then compared to the cluster-analysis results.

The cluster analyses done for this study were run on a computer program (CLUST) written in FORTRAN WATFIV for an IBM3032 system (Savarese, 1984). CLUST is designed to do either Q-or R-mode analysis using a variety of different similarity indices for both binary and numerical-data (for details concerning the clustering technique and the use of various similarity indices see: Valentine and Peddicord, 1967; Fox, 1968; Mello and Buzas, 1968; Cheetham and Hazel, 1969; Farris, 1969; Stephenson and others, 1972; Sneath and Sokal, 1973; Clifford and Stephenson, 1975; MacDonald, 1975). The program utilizes the weighted pair group averaging technique. The subroutine used to print the dendrograms has been borrowed and altered slightly from Davis (1973).

Cluster analysis was used in this study for two reasons: (1) Q-mode and R-mode clustering results could be compared directly to qualitatively observed fossil associations, and (2) Q-mode analysis, by quantitatively defining biofacies, could be used to rigorously test for the existence of faunal cyclicity through the Centerfield strata. Analysis was attempted with two sets of data. First, clustering was done using presence-absence data. All sampling intervals could be included while using presence-absence data, despite the unavailability of bulk-sample data from the limestone beds. The Jaccard index was chosen for calculating similarity levels over the other binary coefficients because this coefficient ignores negative matches between samples (Valentine, 1973). Because the change in fauna appears to be gradational through the sampled intervals and because many of the species occur within several associations with only changes in their absolute-abundance being diagnostic, clustering was repeated using actual numerical abundance data available for a limited subset of samples. Although this method may yield a more precise grouping for the samples included, samples which were not bulk sampled had to be excluded. The 'Canberra-metric' similarity coefficient was chosen for numerical-data clustering because it is accurate for populations with highly variable numbers (Stephenson and others, 1972).

The original presence-absence data matrix appears in Table 1. Thirty-one samples were included (NH = 31). The six difficult to sample intervals were combined into three larger intervals for clustering (22, 23; 24, 25; 26, 27). Not all recorded taxa were included in the clustering. Taxa which appeared very rarely in only one or two samples and those that were found in most samples were excluded, based on criteria outlined by Stephenson and others (1972). Encrusting epibionts and species with questionable identifications were not considered in the data base. After these deletions were made 45 taxa remained for clustering (NS = 45).

The original numerical-data matrix appears in Table 2. Criteria governing reductions in the number of taxa and samples in the standardized original numerical-data matrix are more rigorous for this case. Any interval for which bulk-sample data were not available could not be included in this analysis (intervals 9, 11, 13, 19, 20, 22-27 were removed), leaving a total of 23 included samples (NH = 23). Procedures for removing species from the presence-absence data matrix noted above were also followed here. In addition, easily fragmented fossils (mostly fenestrate bryozoans), and species with uniform numerical distributions throughout (a few brachiopods and bivalves) were deleted, again following procedures of Stephenson and others (1972). A number of species within genera were lumped due to problems in original identification (i.e., *Rhipidomella* and *Paleoneilo* species). These adjustments left a total of 40 taxa (NS = 40).

QUANTITATIVE FAUNAL ANALYSIS

Q-Mode Clustering Results

Q-mode cluster-analysis was run for both the presence-absence and the numerical-data matrices. The two resultant dendrograms appear in Figures 5 and 6. Groupings of samples recognized by this technique are herein termed "biofacies" following the usage of McGhee and Sutton (1981, p. 36); we believe that this term is more appropriate for stratigraphic samples than the commonly used term "biotope" (Kaesler, 1969; Valentine and Peddicord, 1967).

Three very general biofacies (sampled intervals exhibiting similar ecologic conditions as reflected by the fauna) cluster out on the dendrogram produced by the presence-absence data (Fig. 5) and their relative positions are plotted along the composite stratigraphic section in Figure 7.

The biofacies are as follows:

1. Biofacies A groups samples 1-4 and 32-34 at the two extremes of the section. This biofacies corresponds well with the dark-gray clayshale/claystone lithologies. Two smaller subgroups cluster at slightly higher similarity levels within A. The lower four samples remain distinct from the upper three samples at this similarity level and this probably reflects slight faunal differences at the two opposite ends of the stratigraphic section.
2. Biofacies B groups many samples from the medium- to light-gray claystone lithologies. Samples from both above (28-31; 22-25) and below (9; 5-7) the central limestone (sample 19) appear here but, as with biofacies A, each remains as a distinct subgroup at slightly higher similarity levels.

TAXA	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21	22 23	24 25	26 27	28	29	30	31	32	33	34
BRACHIOPODS																															
1. Ambocoelia	1	1	1	1	1	1	1	1	1	1	0	1	0	1	1	0	1	1	0	1	0	1	1	0	1	1	1	1	1	1	1
2. Devonochonetes scitulus	1	1	1	1	1	1	1	1	1	1	0	0	0	0	0	0	0	0	0	0	1	0	0	1	1	1	1	1	1	1	1
3. Leiorhynchus	1	1	1	1	0	0	1	1	0	0	0	1	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	1	1
4. Longispina	1	1	1	1	1	1	1	1	1	1	1	1	1	1	1	1	1	1	1	0	1	0	1	0	1	1	1	0	0	1	0
5. Echinocoelia	1	1	1	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
6. Athyris	0	0	0	0	0	0	0	1	0	1	1	0	1	1	1	0	1	1	1	0	1	1	1	1	1	0	0	0	0	0	0
7. Cyrtina	0	0	0	0	1	0	1	1	1	1	0	1	1	0	1	1	1	1	1	1	1	0	0	0	0	1	0	0	0	0	0
8. Douvillina	0	0	0	0	0	0	0	1	1	1	1	1	0	1	1	1	1	1	1	1	1	0	0	0	0	1	1	0	0	0	0
9. Ilita	0	0	0	0	0	0	1	1	1	1	0	1	0	1	1	1	1	1	1	1	1	0	1	0	1	0	1	0	0	0	0
10. Mediospirifer	0	0	0	0	0	1	1	0	0	1	1	1	1	1	1	1	1	1	1	1	1	1	1	1	0	1	0	0	0	0	0
11. Meristella	0	0	0	0	0	0	0	1	0	0	1	1	1	1	1	1	1	1	1	0	1	1	1	1	0	0	1	0	0	0	0
12. Pholidostrophia	0	0	0	0	0	1	1	1	1	1	1	1	1	0	0	1	0	0	1	0	0	0	0	0	1	0	0	0	0	0	0
13. Parazyga	0	0	0	0	0	0	1	0	0	1	0	0	0	1	1	1	1	1	0	0	1	0	0	0	0	0	0	0	0	0	0
14. Mucrospirifer mucronatus	0	1	1	1	1	1	1	1	0	1	0	1	1	0	0	0	0	0	0	0	1	0	0	0	0	1	1	1	0	1	1
15. Nucleospira	0	1	0	0	1	1	0	1	1	1	0	0	0	1	1	1	1	1	1	0	1	1	1	0	1	1	1	1	0	0	0
16. Productella	0	0	0	0	1	1	0	1	0	1	1	1	1	1	0	0	0	1	0	0	0	0	0	0	1	1	0	0	0	0	0
17. Protoleptostrophia	0	0	0	0	1	1	1	1	0	0	0	0	0	0	0	1	1	1	1	0	1	0	1	1	1	1	1	0	0	0	0
18. Pseudoatrypa	0	0	0	0	0	1	1	1	1	1	1	1	1	1	1	1	0	0	1	0	1	1	1	1	1	1	1	1	1	1	0
19. Rhipidomella	1	1	1	1	1	0	1	1	0	1	1	1	1	1	1	1	1	1	1	0	1	1	0	0	1	1	1	0	0	0	0
CORALS																															
20. Pleurodictyum	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	1	1	0	1	1	0
21. Stereolasma	0	0	0	0	0	0	1	1	0	1	1	1	0	0	1	0	1	1	0	1	1	1	1	1	1	1	1	1	0	0	0
22. Amplexiphyllum	0	0	0	0	0	0	0	0	0	1	0	1	1	0	0	0	0	0	0	0	0	1	0	0	0	1	0	0	0	0	0
23. Heliophyllum	0	0	0	0	0	0	0	0	0	1	1	1	1	1	1	1	1	1	1	1	1	0	0	0	0	0	0	0	0	0	0
24. Eridophyllum	0	0	0	0	0	0	0	0	0	0	0	0	1	1	1	0	1	1	1	1	0	0	0	0	0	0	0	0	0	0	0
25. Cladopora	0	0	0	0	0	0	0	1	0	0	0	0	1	1	1	1	1	1	1	0	1	0	0	0	0	1	0	0	0	0	0
26. Biotherophyllum	0	0	0	0	0	0	0	0	0	0	0	1	0	1	1	1	1	1	1	1	0	0	0	0	0	0	0	0	0	0	0
27. Aulocystis	0	0	0	0	1	1	1	1	1	1	0	1	1	1	1	1	1	1	1	1	1	1	1	0	1	1	1	1	1	1	1
28. Favosites sp.	0	0	0	0	0	0	1	1	0	1	1	1	1	1	1	1	1	1	1	1	1	0	0	0	0	0	0	0	0	0	0
BIVALVES																															
29. Actinopteria	0	1	1	0	0	1	0	1	1	1	1	1	1	1	0	0	1	1	1	0	1	0	0	0	0	0	0	0	0	0	0
30. Cypricardinia	0	0	0	0	1	1	1	1	0	1	0	0	1	1	0	1	1	1	0	0	0	1	1	0	1	1	1	1	1	0	1
31. Modiomorpha sp.	0	1	1	0	0	0	0	1	0	0	0	0	1	0	0	0	0	1	0	0	0	0	0	0	0	0	0	0	0	0	0
32. Nuculoidea	1	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	1
33. Nuculites	0	1	1	1	0	1	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	1	0	0	0	1	0
34. Pterochaenia	1	1	1	1	0	0	0	0	0	1	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	1	1	1	1	1	1
GASTROPODS																															
35. Bucanopsis	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	1	1	1
36. Naticonema	0	0	0	1	0	0	1	1	1	1	1	1	1	1	0	0	0	0	0	0	1	0	1	0	0	0	1	0	0	0	0
37. Paleozygapleura	0	1	0	0	1	0	1	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	1	1	1
38. Platyceras	0	0	0	0	0	0	0	1	0	1	1	1	1	0	1	1	1	1	1	0	1	0	0	0	0	1	0	1	0	0	0
TRILOBITES																															
39. Greenops	1	1	1	0	1	1	1	1	0	1	1	1	0	0	0	0	0	1	0	0	1	0	1	0	1	1	1	1	1	1	1
40. Phacops	0	0	0	1	1	1	1	1	1	1	1	1	1	1	0	1	1	1	1	1	1	1	1	1	1	1	1	1	1	1	1
BRYOZOANS																															
41. Fenestella eumaciata	0	0	0	0	0	0	1	1	1	1	1	1	1	1	1	1	1	1	1	1	1	1	1	0	1	1	1	1	0	0	0
42. F. planiramosa	0	0	0	0	1	0	0	0	0	0	0	0	1	0	0	0	0	0	0	0	1	1	1	0	1	1	1	1	0	0	0
43. fistuliporid	0	0	0	0	0	0	1	1	1	1	1	1	1	1	1	1	1	0	1	1	1	1	1	0	1	1	1	1	0	0	0
44. Loculipora	0	0	0	0	0	0	0	1	0	1	1	1	1	0	1	1	1	1	1	1	1	1	1	0	1	1	1	1	0	0	0
45. Sulcoretepora	0	0	0	0	1	1	1	1	1	1	1	1	1	1	1	0	1	1	0	1	1	0	1	0	1	1	1	1	0	0	0

Table 1: Original presence-absence data for 45 taxa in all 34 horizons. Horizons 22, 23 & 24, 25 & 26, 27 were combined. A 'O' denotes an organism's absence, a '1' its presence.

3. Biofacies C includes all but two of the remaining samples. 'C' is concentrated at the center of the section and corresponds to the argillaceous limestone and calcareous light-gray claystone lithologies. Two very distinct subgroups are defined at high levels and are, therefore, worth noting and differentiating. Subgroup C2 includes the most centrally located samples (14-20), while subgroup C1 appears immediately below and above C2 (Fig. 7).

Q-mode clustering using the numerical-data yielded similar results. Although large portions of the section could not be included, the finer cluster resolution available due to the nature of the data produced a more refined subdivision of biofacies. Five groups are recognizable from the dendrogram (Fig. 6) and their temporal distribution through the Centerfield section is plotted on Figure 7.

The biofacies are as follows:

1. Biofacies 'a' groups the four upper samples (31-34) and the four lower samples (1-4). Biofacies 'a' is found associated with the dark-gray clayshale/claystone lithologies and its position corresponds well with that of biofacies A of the presence-absence data clustering.

TAXA	SAMPLING HORIZONS																						
	1	2	3	4	5	6	7	8	10	12	14	15	16	17	18	21	28	29	30	31	32	33	34
BRACHIOPODS																							
1. Ambocoelia	30	136	181	92	168	133	367	106	7	20	5	1	0	1	2	0	1	9	5	5	11	8	5
2. Devonochonetes scitulus	21	28	40	37	39	74	35	4	5	0	0	0	0	0	0	2	7	8	7	318	166	89	347
3. Leiorhynchus	119	188	31	1	0	0	1	2	0	3	0	0	0	0	0	0	0	0	0	0	0	1	2
4. Longispina	11	18	39	71	41	132	176	101	25	27	8	6	4	5	6	2	9	4	6	0	0	1	0
5. Echinocoelia	192	32	145	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
6. Athyris	0	0	0	0	0	0	0	2	1	0	1	2	0	9	10	5	1	0	0	0	0	0	0
7. Cyrtina	0	0	0	0	1	0	1	11	21	21	0	1	4	4	7	2	0	1	0	0	0	0	0
8. Douvillina	0	0	0	0	0	0	0	1	3	8	2	1	1	2	3	25	0	2	2	0	0	0	0
9. Elita	0	0	0	0	0	0	1	1	5	9	1	2	7	9	16	5	1	0	1	0	0	0	0
10. Mediospirifer	0	0	0	0	0	2	1	0	1	2	2	3	2	6	13	8	0	2	0	0	0	0	0
11. Meristella	0	0	0	0	0	0	0	1	0	7	3	2	2	1	17	5	0	0	1	0	0	0	0
12. Pholidostrophia	0	0	0	0	0	2	10	32	45	8	0	0	1	0	0	0	1	0	0	0	0	0	0
13. Parazyga	0	0	0	0	0	0	11	0	1	0	2	2	0	5	6	1	0	0	0	0	0	0	0
14. Mucrospirifer mucronatus	0	5	7	1	3	3	2	3	1	5	0	0	0	0	0	1	0	2	1	2	0	2	7
15. Nucleospira	0	5	0	0	1	1	0	19	8	0	4	6	3	6	18	7	5	2	1	2	0	0	0
16. Productella	0	0	0	0	1	4	0	17	1	5	2	0	0	0	1	0	2	7	0	0	0	0	0
17. Protoleptostrophia	0	0	0	0	1	1	8	4	0	0	0	0	1	1	2	1	2	2	3	0	0	0	0
18. Pseudoatrypa	0	0	0	0	0	5	4	18	10	38	14	10	6	0	0	4	172	79	36	4	1	1	0
19. Rhipidomella	2	11	5	1	2	0	3	3	9	28	2	2	7	17	1	18	6	5	1	0	0	0	0
CORALS																							
20. Pleurodictyum	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	1	1	0	1	2	0
21. Stereolasma	0	0	0	0	0	0	1	3	8	5	0	1	0	4	0	0	16	17	10	2	0	0	0
22. Amplexiphyllum	0	0	0	0	0	0	0	0	4	3	0	0	0	0	0	0	0	1	0	0	0	0	0
23. Heliophyllum	0	0	0	0	0	0	0	0	6	17	22	22	24	12	38	4	0	0	0	0	0	0	0
24. Eridophyllum	0	0	0	0	0	0	0	0	0	0	2	1	0	4	7	0	0	0	0	0	0	0	0
25. Cladopora	0	0	0	0	0	0	0	1	0	0	27	9	9	8	7	2	0	1	0	0	0	0	0
26. Biotherophyllum	0	0	0	0	0	0	0	0	0	1	13	18	7	20	31	0	0	0	0	0	0	0	0
27. Aulocystis	0	0	0	0	5	8	7	18	14	23	99	21	15	28	55	19	38	142	148	114	1	32	5
28. Favosites sp.	0	0	0	0	0	0	2	1	1	4	129	75	15	44	93	6	0	0	0	0	0	0	0
BIVALVES																							
29. Actinopteria	0	2	1	0	0	1	0	3	3	9	2	0	0	1	2	1	0	0	0	0	0	0	0
30. Cypricardinia	0	0	0	0	1	2	4	3	3	0	4	0	1	2	3	0	1	7	14	3	1	0	1
31. Modiomorpha sp.	0	8	15	0	0	0	0	1	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
32. Nuculoidea	19	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	4
33. Nuculites	0	13	4	2	0	1	0	0	0	0	0	0	0	0	0	0	0	1	0	0	0	1	0
34. Paleoneilo sp.	0	0	3	0	0	3	0	0	0	0	0	0	0	0	0	2	1	2	2	0	0	0	0
35. Pterochaenia	21	5	2	2	0	0	0	0	1	0	0	0	0	0	0	0	0	5	1	14	14	3	5
GASTROPODS																							
36. Naticonema	0	0	0	1	0	0	1	3	1	1	1	0	0	0	0	6	0	0	1	0	0	0	0
37. Paleozygapleura	0	4	0	0	1	0	1	0	0	0	0	0	0	0	0	0	0	0	0	0	5	8	9
38. Platyceras	0	0	0	0	0	0	0	5	6	8	0	1	1	2	1	17	0	1	0	1	0	0	0
TRILOBITES																							
39. Greenops	1	12	1	0	1	5	18	9	7	5	0	0	0	0	1	1	1	1	1	10	5	3	14
40. Phacops	0	0	0	1	7	15	18	17	10	8	2	0	2	2	4	35	20	9	2	1	1	1	2

Table 2: Original standardized numerical data for 40 taxa in a subset of 23 horizons.

2. Biofacies 'b' is found associated with the gray claystone lithology in the samples located just within the the position of biofacies 'a' (28-30; 5-6).
3. Biofacies 'c' and 'd' are both small groups, each containing two samples (7 and 8 for 'c'; 10 and 12 for 'd'). These two biofacies are found associated with the lower light-gray claystones and do not appear in the upper part of the section, probably because of incomplete sampling.
4. Biofacies 'e' includes the samples adjacent to the middle limestone (14-18). Sample 21 also was included in this biofacies because it was more similar to this group than any other. The group is found associated with the lightest gray, most fossiliferous claystones and corresponds well with the position of subgroup C1 resulting from the clustering with presence-absence data.

R-Mode Clustering Results

R-mode cluster-analysis was similarly run for both the presence-absence and numerical-data matrices. R-mode results were not as significant as the Q-mode results. These dendrograms show considerably more scatter and groups cluster at lower similarity levels. Only two groups produced for the numerical-data (Fig. 6), for example, are defined at a similarity level of approximately 0.58. Movement of this phenon line to higher similarity values produces a large number of small, insignificant groups. These groups also do not compare well with the semiquantitatively recognized faunal associations described below.

The two-way diagram (Fig. 6) can be used to help understand these results and some of the problems associated with this

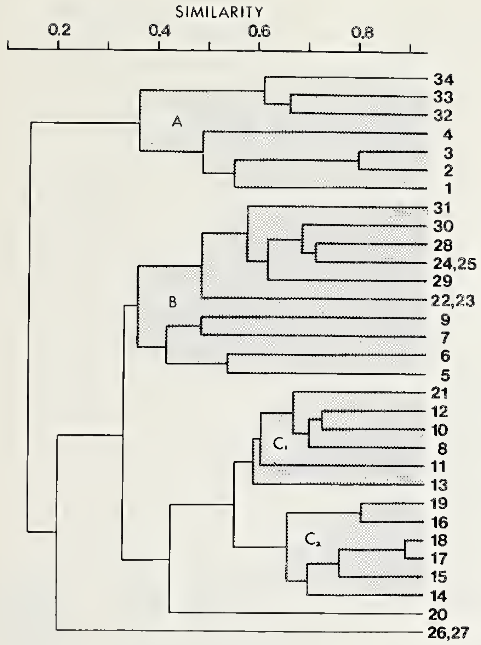


Figure 5: Q-mode cluster analysis dendrogram for the presence/absence data matrix. The Jaccard similarity levels are listed on the abscissa and the 34 sampled horizons are on the ordinate (note that a few horizons were combined; i.e., 24 and 25). Three biofacies are defined (A,B,C) with two distinct subgroups in biofacies C.

particular clustering technique. The Q- and R-mode dendrograms for the numerical-data are plotted in this figure. Species 1-40 and samples 1-34 appear on the abscissa and ordinate, respectively, of the grid. A darkened square denotes the presence of any given species within a particular sample. The five numerically defined biofacies are shaded on the Q-mode dendrograms. Two general biofacies (also shaded) defined along with many species clustering at low, insignificant levels appear on the R-mode dendrogram. A pattern in species distribution is evident when studying the darkened regions of this figure. Species tightly linked within the two general biofacies are restricted to a few sampling intervals or a few biofacies. The species clustering at insignificant levels on the right are very ubiquitous and occur throughout many samples, within all biofacies. This clustering technique, with its two-dimensional discrete nature, can only consider species once while linkages are made. The ubiquitous species can only appear in one group and, therefore, cluster at very low levels where they are linked with many more species (see McGhee and Sutton, 1981, for a discussion of similar R-mode results).

Faunal Patterns

A distinct faunal cycle paralleling the lithologic cycle is evident from the distribution of biofacies derived from the binary-data results (Fig. 7). With two minor exceptions (the clustering of samples 26 and 27 at insignificant levels and the appearance of sample 8 in biofacies C), the distribution is completely symmetrical, with subgroup C2 (qualitatively characterized by *Favosites hamiltoniae*) marking a center of symmetry.

A similar cyclic distribution pattern is seen for the numerically derived biofacies clustering results (Fig. 7). Biofacies 'a, b, and e' are located symmetrically about the Centerfield middle limestone. The disruption in the pattern, due to the non-appearance of biofacies 'c' and 'd' above the middle limestone, can be explained by the absence of absolute-abundance data from samples 22-27 of the original data set. Had data from these samples been available and included, the symmetry probably would have been complete. Regardless of the cause, a cyclic change in biofacies and, by implication, of environmental setting must have occurred through the time of deposition of the complete Centerfield sequence.

Quantitative analysis also reveals that the upper and lower beds of the Centerfield form a continuous gradation. With one exception, there are no reversals in sequence above or below the middle limestone. Thus, for example, there are no anomalous occurrences of biofacies A (characterized by chonetid and *Ambocoelia* brachiopods) adjacent to the middle limestone; conversely, no samples of biofacies C are inserted among the transitional medium-gray claystones on either side of the middle. Rather, the sequence is regular; the sequence of biofacies above the middle limestones is essentially a mirror image of the sequence below those beds.

Overall patterns of faunal richness, equitability, and the relative abundances of species throughout the section further corroborate the cyclic nature of the depositional history. The Shannon-Weaver dominance diversity index was calculated for most of the sampling intervals (see Table 3; diversities could not be determined for poorly exposed portions of the section [samples 20, 21-27] and the limestones). Diversities were then ranked relatively, and these diversity grades are plotted next to the composite section (Fig. 7).

An interesting pattern emerges. Despite the gaps along the section, a cyclic pattern in diversity is evident, paralleling the change in biofacies. Diversity levels are low at the base and top of the section, corresponding to biofacies A (*Leiorhynchus* — styliolinid and *Ambocoelia* — chonetid associations, see below). Diversity levels increase progressively inward toward the middle limestones. A high level of diversity can also be estimated for the middle limestones based on species richness as determined from field counts (K = 36 for sample 19; Table 1).

Patterns of equitability can also be investigated for a similar cyclicity (Table 1). These values were ranked similarly and these ranked values also appear along the composite section (Fig. 7). A similar low-high-low pattern is apparent.

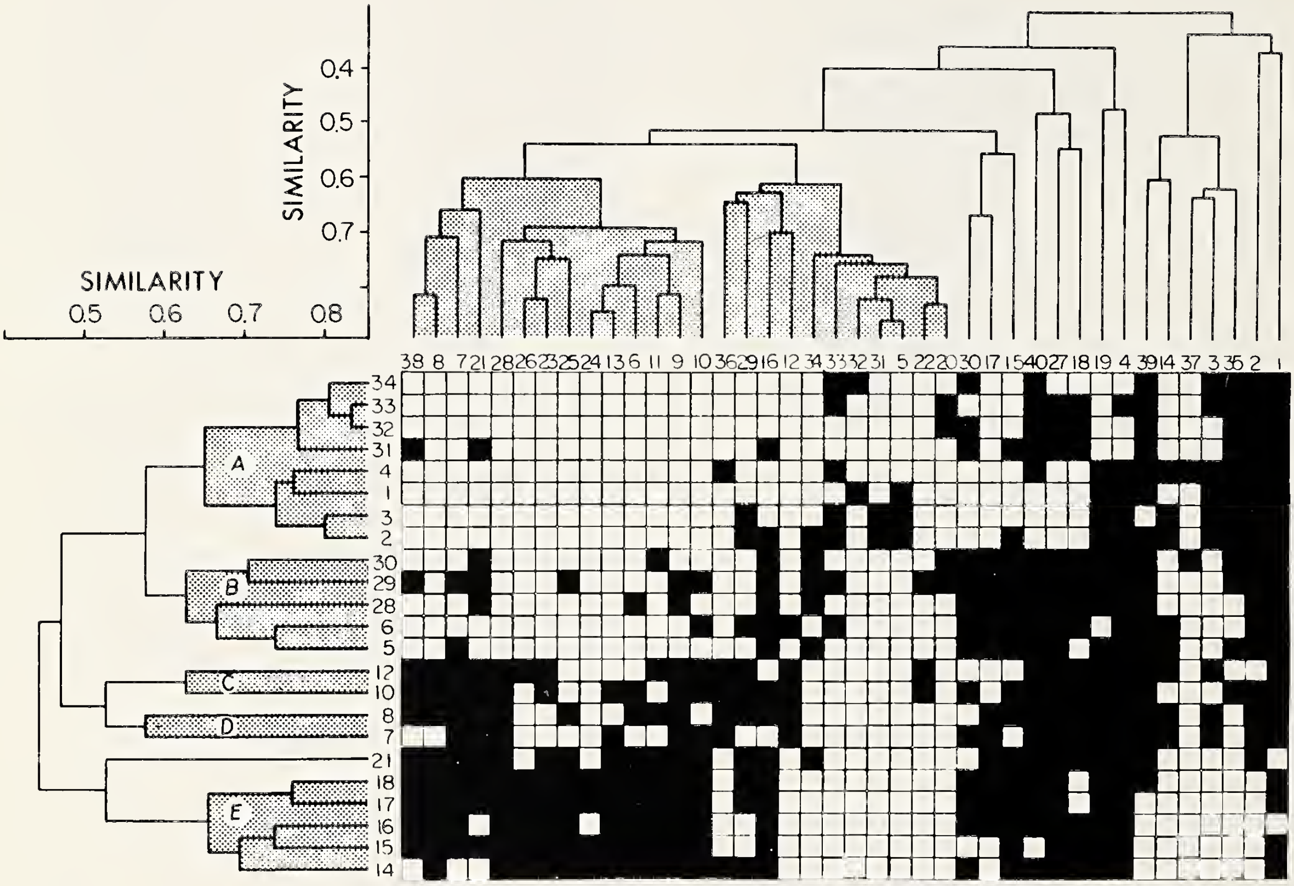


Figure 6: A two way diagram, plotting Q- vs. R-mode resultant dendrograms from the numerical data. The 34 sampled horizons and biofacies a through e appear on the ordinate and the 40 taxa appear on the abscissa. A darkened box within the gridwork appears when a particular organism is found in that horizon. Organisms represented by numbers 1–40 can be found in table 2.

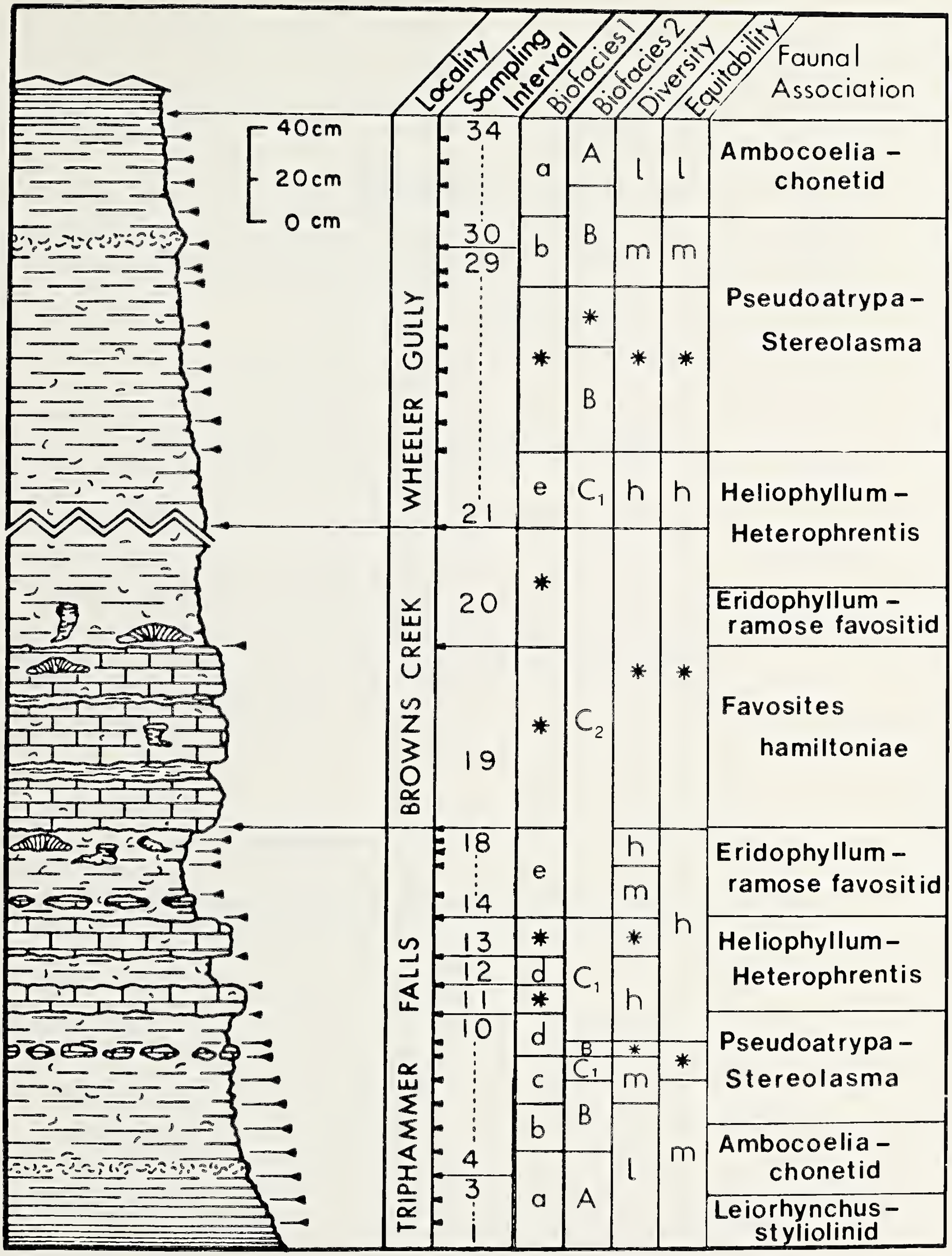


Figure 7: Centerfield composite section with the Q-mode cluster analysis biofacies and corresponding faunal association names. Biofacies 1, biofacies from the numerical data; biofacies 2, biofacies from the presence/absence data. Asterisks mark portions of the section that could not be included in the clustering. Shannon/Weaver diversity and equitability grades also appear in the last two columns. For diversity: l = low (0. 30 – 0.70); m = medium (0.70 – 1.20); h = high (greater than 1.20). For equitability: l = low (0.00 – 0.50); m = medium (0.50 – 0.75); h = high (0.75 – 1.00). Here asterisks denote horizons for which numerical data were lacking and, hence, indices could not be calculated.

sampling interval	brachiopods	corals	bryozoans	bivalves	gastropods	trilobites	others	K	N	H'	E
34	9	1	0	5	2	2	0	19	546	0.3533	0.2763
33	9	2	0	4	3	2	0	20	217	0.7116	0.5470
32	8	2	0	3	3	2	3	21	289	0.4824	0.3648
31	8	2	7	4	2	2	4	29	678	0.5851	0.4001
30	18	4	7	7	2	2	6	46	472	1.0464	0.6293
29	19	4	8	7	1	2	6	47	711	1.1788	0.7050
28	16	2	7	2	1	2	1	31	486	0.8838	0.5926
27	–	–	–	–	–	–	–	–	–	–	–
26	–	–	–	–	–	–	–	–	–	–	–
25	–	–	–	–	–	–	–	–	–	–	–
24	–	–	–	–	–	–	–	–	–	–	–
23	–	–	–	–	–	–	–	–	–	–	–
22	–	–	–	–	–	–	–	–	–	–	–
21	22	12	10	3	4	2	2	55	419	1.4445	0.8300
20	6	14	4	1	0	1	0	26	253	–	–
19	21	9	3	1	1	1	0	36	210	–	–
18	26	15	3	2	1	2	3	52	394	1.2912	0.7524
17	16	11	4	2	1	1	2	37	233	1.3532	0.8629
16	19	7	4	1	1	1	3	36	131	1.4045	0.9025
15	19	11	6	0	1	0	2	39	231	1.1742	0.7380
14	20	9	4	3	1	1	2	40	397	1.0746	0.6708
13	15	13	8	4	3	2	2	47	244	–	–
12	20	6	5	3	2	3	3	42	384	1.3803	0.8503
11	–	–	–	–	–	–	–	–	–	–	–
10	22	5	6	6	4	2	2	47	322	1.3282	0.7943
9	11	1	5	3	1	1	0	22	61	–	–
8	22	3	6	9	4	3	3	50	448	1.1946	0.7031
7	15	3	3	3	2	2	1	29	692	0.7198	0.4922
6	14	1	1	10	0	2	2	30	432	0.8571	0.5803
5	9	1	1	3	1	2	2	19	280	0.6061	0.4740
4	7	0	0	3	1	1	1	13	216	0.6014	0.5399
3	9	0	0	5	0	1	3	18	357	0.7424	0.5914
2	10	0	0	6	2	1	1	20	594	0.8206	0.6307
1	7	0	0	3	0	1	2	13	450	0.7481	0.6716

Table 3: Number of species found for each fossil group, total number of species (K), total number of fossil individuals (N), the Shannon/Weaver diversity (H') and equitability (e) indices for each of the 34 sampled intervals. Dash entries were made when data were not available.

The two spindle diagrams show the ranges and relative abundances of the most important coral (Fig. 8) and brachiopod taxa (Fig. 9). Low diversity, low equitability assemblages found at the extremes of the composite section are dominated by a few species with high relative abundances. The more diverse, equitable assemblages concentrated near the center of the section are characterized by many taxa, each with lower relative abundances. Based on this Centerfield section and other Hamilton units, taxa can be categorized as stenotypic (found in single biofacies) or eurytypic (found in several biofacies). Note that certain taxa are restricted to particular portions of the sequence, whereas others occur in most units.

FAUNAL ASSOCIATIONS

Based on results of cluster analysis and field observations, a total of five faunal associations can be recognized. Although clusters correspond approximately to qualitatively discernable fossil distributions, problems inherent to the clustering technique do not allow for as refined a subdivision of fossil associations (see McGhee and Sutton, 1981). Therefore, the following associations have been defined using a combination of quantitative and observational data. This permits greater subdivision of paleoenvironments. The following sections provide detailed descriptions of each association including a discussion of characteristic taxa, corresponding lithology, and sedimentologic and taphonomic features. Each description is followed by a brief discussion of the inferred depositional setting.

Association 1: *Leiorhynchus* — Styliolinid

The brachiopods *Leiorhynchus limitare, Devonochonetes scitulus*, and *Echinocoela* sp.are the most numerous fossils, all having greater than 10% abundances (Fig. 9). Styliolinids and plant spores (?*Tasmanites*) also are very abundant. Other common brachiopods include: *Lingula delia, Ambocoelia umbonata*, and *Pholidops*. Shallow infaunal clams (e.g., *Paleoneilo, Modiomorpha*, nuculids) as well as occasional nautiloids are present; corals are absent. Diversity levels are low for samples containing this association (see Table 1 for samples 1-2; Fig. 7), yet the abundances for those species which are present are very high (Fig. 9).

This association is found in dark-gray to black fissile, brittle shales of the uppermost Levanna Shale Member, low on the composite section (samples 1-2; Fig. 7). The shales are rich in organics and emit a petroliferous odor when split. The same association with slight modifications, most notably the absence of *Echinocoelia*, can be found above the Centerfield within the middle Ledyard Shale; however, this sample was not included in the quantitative sampling.

Cluster analysis grouped this association within biofacies 'a' for the numerical-data results and biofacies 'A' for the binary-data results (Fig. 7).

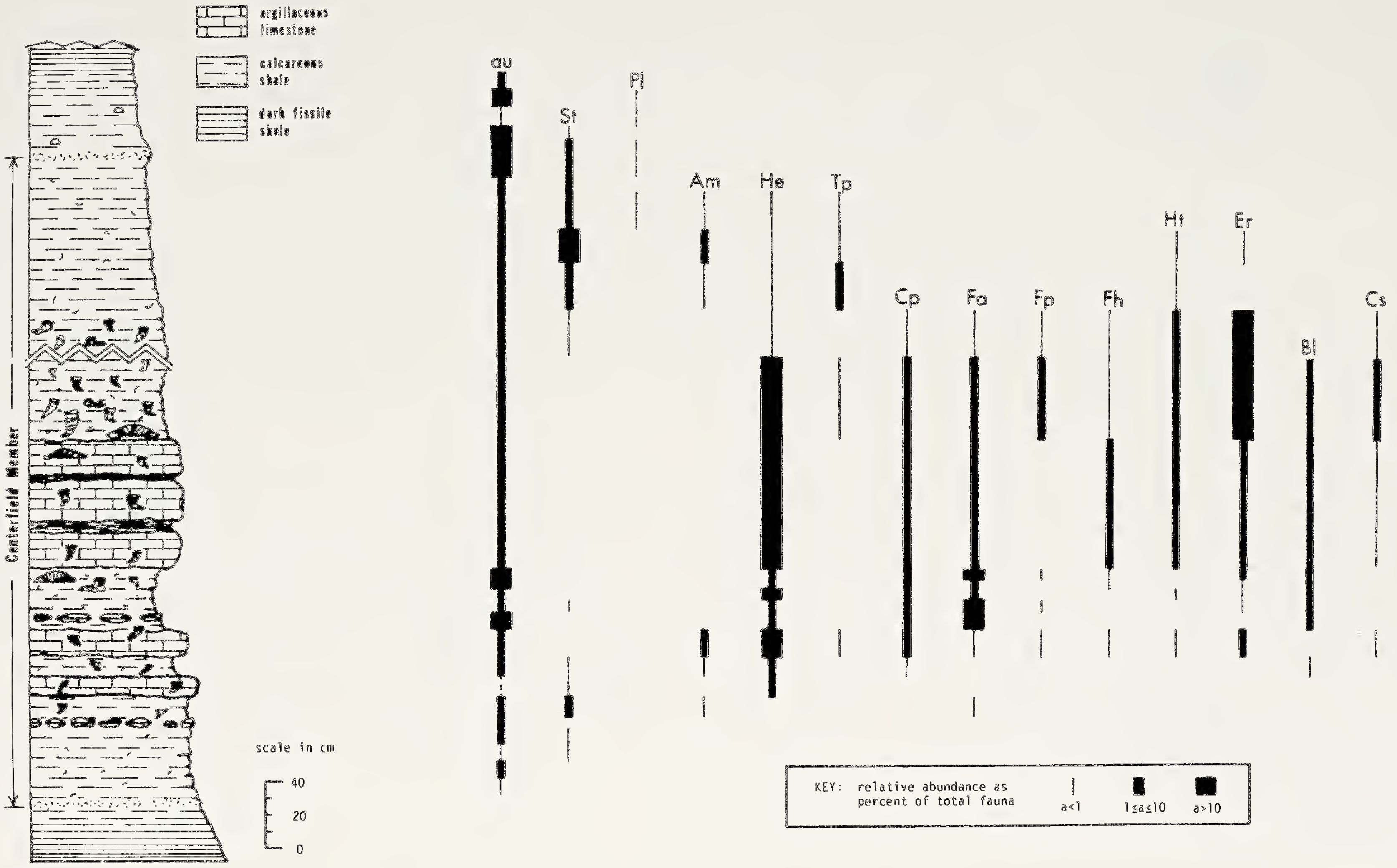


Figure 8: Spindle diagram showing the range and relative abundances of the important coral constituents throughout the Centerfield composite section. Length of the line depicts the range of occurrence and the thickness of the line represents the relative abundance. Au, auloporids (mostly *Aulocystis*); St, *Stereolasma*; Pl, *Pleurodictyum*; Am, *Amplexiphyllum*; He, *Heliophyllum*; Tp, *Thamnoptychia*; Cp, *Cladopora*; Fa, *Favosites arbuscula*; Fp, *F. placenta*; Fh, *F. hamiltoniae*; Ht, *Heterophrentis*; Er, *Eridophyllum*; Bl, *Blothrophyllum*; Cs, *Cystiphylloides*.

The abundant *Leiorhynchus limitare* valves commonly are fragmented but this is probably due, in part, to compaction of these thin-shelled brachiopods. When found here unfragmented, they are strongly compressed and highly fractured. Fragmentation may have been biogenic or could have resulted from impacts of slightly decalcified shells during episodic current agitation. The styliolinid and orthoconic nautiloids display vague preferred orientations indicative of weak bottom currents. Other fossils usually are not fragmented; brachiopods such as *Echinocoelia* are generally disarticulated. Chonetids frequently have intact hinge spines. The occasional *Greenops boothi* occur as either complete or molt specimens. Fossils are not encrusted or corroded, in marked contrast to shells obtained from other associations.

The high degree of disarticulation and fragmentation of shells from this association points to generally very low rates of sedimentation; the seafloor was apparently disturbed by very gentle current action. The dark color along with the high organic content of the shale indicates a low-energy, dysaerobic setting for this association.

Shale fissility and preservation of oriented styliolinid fabrics indicate a general lack of bioturbation; few trace fossils are found associated with this lithology. These observations point to anoxic conditions in the sediment. However, the occurrence of nuculid and modiomorphoid bivalves on certain bedding planes suggests fluctuating oxygen levels near the sediment water interface. Nuculoids and articulate brachiopods are known to tolerate very low oxygen concentrations (Rhoads and Morse, 1971; Thayer, 1981).

Association 2: *Ambocoelia* — Chonetid

The chonetid brachiopods *Longispina mucronatus* and *Devonochonetes scitulus* and the small spiriferid *Ambocoelia umbonata* are the most abundant fossils present in this association (Fig. 9). Other brachiopods, e.g., *Mucrospirifer mucronatus* and *Schuchertella*, are common but subordinate members. As the lithology grades to a lighter gray claystone, the assemblage also may include rare fenestrate bryozoans. The ramose auloporid coral *Aulocystis jacksoni* (formerly assigned to *Cladochonus*, Laub, 1972) is present to common (see Fig. 9, for definitions). (The corresponding coral association previously has been referred to as the *Aulocystis* association; see Brett et al., 1983.) Other common taxa include the trilobites *Greenops* and *Phacops*, the gastropods *Mourlonia* and *Naticonema*, and several genera of small bivalves. Unlike the *Leiorhynchus* — styliolinid assemblage, trace fossils are present to common in occurrence and include pyritized vertical tubes, thread-like horizontal burrows filled with limonite and small *Zoophycos*. Diversity levels are intermediate between those of the *Leiorhynchus* — styliolinid and *Pseudoatrypa* — *Stereolasma* associations (see Fig. 7 and Table 1 for samples 3-5, 31-34).

Although the same general faunal constituents are present both near the base and at the top of the Centerfield section (samples 3-5 and 31-34), some differences above and below are discernable. The chonetid *Longispina mucronatus* is less common and *Devonochonetes scitulus* is more common in samples 31-34 than in samples 3-5. *Aulocystis jacksoni* is much more common above and the tabulate coral *Pleurodictyum americanum* is present above but absent below.

This association is found in the dark-to medium-gray, soft, friable clayshale to claystone lithologies. These lithologies are found low in the Centerfield section (samples 3-5), directly overlying the dark-gray shales of the *Leiorhynchus* — styliolinid association and reappear higher in the section (samples 31-34), just below the upper dark gray clayshales of the Ledyard (Fig. 7). This association clusters as the major component of biofacies 'a' for the numerical-data results and as the major component of biofacies 'A' for the binary-data results (Fig. 10).

Brachiopod shells typically are disarticulated, with articulation-disarticulation ratios for *Mucrospirifer mucronatus* and *Ambocoelia umbonata* being less than one. Shell breakage, however, is slight. *Aulocystis dichotoma* colonies commonly were preserved intact. Trilobite material seen *in situ* consisted mainly of unbroken cephala and pygidia. Rare whole specimens and thoracopygidia (moults?) were found. Infaunal burrowers were present within the bottom sediments at time of deposition. Fissility is poorly developed; pyritized tubules and *Zoophycos* swirls are present.

This faunal association lived in normally low-energy environments with slow to moderate rates of sedimentation, as indicated by the slight breakage and moderate disarticulation of the fossils. The dark-gray coloration of the rocks along with the near absence of fissility suggest shifting dysaerobic to aerobic conditions at the sediment-water interface. This assemblage is inferred to have accumulated within the photic zone because it contains trilobites with well developed eyes and probable herbivorous gastropods.

Association 3: *Pseudoatrypa* — *Stereolasma*

This association is characterized by a diverse brachiopod assemblage, with other faunal elements being of lesser importance. The association, however, has a high degree of variability in its brachiopod constituents throughout its occurrence. Three species of brachiopods stand out: *Pseudoatrypa cf. P. devoniana* is common to abundant throughout the association; *Mucrospirifer consobrinus* is common where the association appears above the Centerfield middle limestone (samples 22-30) but is absent below, whereas, *Pholidostrophia nacrea* is present to abundant in the association below the middle limestone (sample 6-11) and is absent above (Fig. 9). A number of other brachiopod genera are of secondary importance, including *Ambocoelia, Cyrtina, Douvillina, Longispina, Mediospirifer, Nucleospira, Productella*, and *Rhipidomella*.

The coral constituents also show slight differences above and below. The rugose corals *Stereolasma rectum* and *Amplexiphyllum* and the tabulate *Aulocystis jacksoni* are ubiqui-

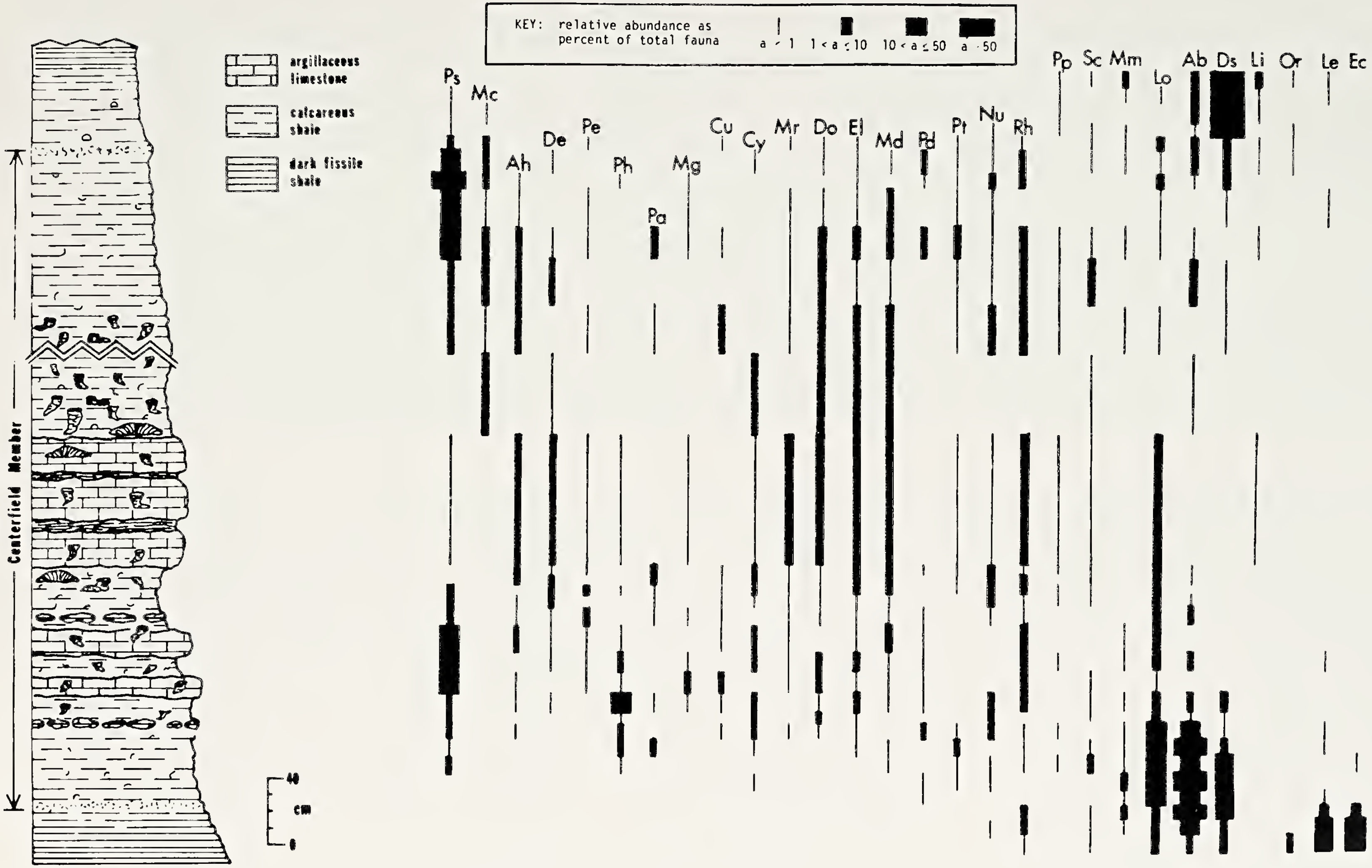


Figure 9: Spindle diagram showing the range and relative abundances of the important brachiopod constituents throughout the Centerfield composite section. Length of the line depicts the range of occurrence and the thickness of the line represents the relative abundance. **Ps,** *Pseudoatrypa;* **Mc,** *Mucrospirifer consobrinus;* **Ah,** *Athyris;* **De,** *Delthyris;* **Pe,** *Pentamerella;* **Ph,** *Pholidostrophia;* **Pa,** *Parazyga;* **Mg,** *Megastrophia;* **Cu,** *Cupulorostrum;* **Cy,** *Cyrtina;* **Mr,** *Meristella;* **Do,** *Douvillina;* **El,** *Elita;* **Md,** *Mediospirifer audaculus;* **Pd,** *Productella;* **Pt,** *Protoleptostrophia;* **Nu,** *Nucleospira;* **Rh,** *Rhipidomella;* **Pp,** *Pholidops;* **Sc,** *Schuchertella;* **Mm,** *Mucrospirifer mucronatus;* **Lo,** *Longispina;* **Ab,** *Ambocoelia;* **Ds,** *Devonochonetes scitulus;* **Li,** *Lingula;* **Or,** *Orbiculoidea;* **Le,** *Leiorhynchus;* **Ec,** *Echinocoelia.*

tous in this association. Other tabulates (*Pleurodictyum* and *Thamnoptychia*), however, are present only in the upper representatives of the association (Fig. 8). Small specimens of the rugose coral *Heliophyllum halli* are present both above and below. Bryozoans (*Fenestella emaciata, Sulcoretepora, Loculipora*), crinoid columnals, platycerid snails, and the trilobites *Greenops* and *Phacops* are also common. *Zoophycos* trace fossils are abundant.

Richness levels are very high. Some of the highest calculated Shannon-Weaver diversity indices occurred in this association (see Fig. 7 and Table 1 for samples 6-10, 22-30). Fossil abundances for each species tended to be lower than in the two previous associations (Figs. 8 and 9) and species occurrence counts were more equitable overall.

Although the *Pseudoatrypa* — *Stereolasma* association occurs both above and below the middle limestone, the association above is represented by a thicker interval (approximately 1 m). Below the middle limestone, the association is confined to approximately 60 cm of section. The most obvious faunal differences between the two is the appearance of the brachiopod *Mucrospirifer consobrinus* and the corals *Pleurodictyum* and *Thamnoptychia* above and not below. These corals may have been more sediment-tolerant than *Stereolasma* and *Amplexiphyllum* and their presence may indicate a higher sedimentation rate for the depositional environments found for this association in the upper submember (Brett and others, 1983).

This association occurs in medium to light-gray, highly bioturbated claystones. This lithofacies is found in the upper half of the composite section just below the appearance of association 2 (samples 22-30 on Fig. 7). It also is found in the lower half of the Centerfield section just above the dark-gray claystones corresponding to the appearance of the *Ambocoelia* — chonetid association and below the lowermost of the two thin-bedded argillaceous limestones (samples 6-10). This association corresponds fairly well with biofacies 'b and d' from the numerical-data clustering results and biofacies 'B' for the binary-data results (Fig. 7).

Brachiopods in this association are almost exclusively disarticulated with moderate amounts of fragmentation. Fenestrate and ramose bryozoan colonies, on the other hand, are highly fragmented here. Trilobites, as well as crinoids, typically are disarticulated. Fossils are typically packed together in chaotic orientations.

The fine grained nature of the sediment indicates generally low-energy conditions. The high degree of disarticulation with moderate fragmentation, however, suggests periods of higher energy levels. The higher diversity together with an increase in the number of infaunal species (e.g., abundant *Zoophycos*) indicate a better oxygenation of the sea floor than in the previous association. This association, therefore, lived within depths of normally quiet-water but within reach of the deepest storm waves.

Association 4: Large Rugose Coral

This association is dominated by large rugose corals including *Heliophyllum, Cystiphylloides, Heterophrentis, Eridophyllum*, and *Blothrophyllum*, and ramose favositids, most notably *Favosites arbuscula, F. milne-edwardsii*, and *F. placenta*. Other corals are present but are of secondary importance, including *Stereolasma, Thamnoptychia, 'Cladopora,' Alveolites, Favosites turbinatus* and *Aulocystis*. Small encrusting stromatoporoids are common. Brachiopods are very diverse, but each species occurs in low relative abundance (see Fig. 9 for samples 11-18, 20-21). The taxa present are very similar to the brachiopods seen in the *Pseudoatrypa* — *Stereolasma* association; common genera include: *Pseudoatrypa, Rhipidomella, Mediospirifer, Elita, Pentamerella, Meristella, Parazyga, Delthryis, Cyrtina, Nucleospira*, and *Longispina*. Fenestellid bryozoan and crinoid debris are common. The trilobites are dominated by *Phacops*.

Two subassociations can be recognized within the large rugose coral association, although these two are completely intergradational. The first, called the *Heliophyllum* — *Heterophrentis* subassociation, is dominated by these large solitary rugosans; *Eridophyllum* and ramose favositids are uncommon. The second, the *Eridophyllum* — ramose favositid subassociation, is characterized by the colonial rugose *Eridophyllum subcaespitosum* and the three ramose favositids previously mentioned, with lesser numbers of the solitary rugosans. The two subassociations are very distinct above the middle limestone but tend to be more intergradational below (Fig. 7). Subtle differences in environmental tolerances of these two subassociations, coupled with slow rates of sedimentation, have probably caused time-averaging in the lower unit.

Diversity levels for the large rugose association, as a whole, are higher than any other association (Fig. 7 and Table 1). Relative abundances are low and equitable for most fossils (Figs. 8 and 9).

The large coral association is found in the light-gray calcareous claystones bordering the argillaceous central limestone and in the lower two thin limestones. The *Heliophyllum* — *Heterophrentis* subassociation is found above in sample 21 and in part of 20 and the *Eridophyllum* — ramose favositid subassociation is found just below in the remainder of sample 20 (Fig. 7). The position of these two is not as discrete below the middle limestone. They are found intergradationally throughout samples 11-18 with *Heliophyllum* and *Heterophrentis* concentrated in samples 11-13. Cluster analysis for the binary-data equates the *Heliophyllum* — *Heterophrentis* subassociation with biofacies 'C1' and the *Eridophyllum* — ramose favositid subassociation with part of biofacies 'C2' (Fig. 7). Clear relationships between this association and the numerical-data results are not evident.

Although the preservational quality for both subassociations is similar, taphonomic features do differ markedly between the claystone and the lower thin argillaceous limestone lithofacies. Within the light-gray calcareous claystones, brachiopods are generally disarticulated (articulation ratio<1) and often highly fragmented. Brachiopods have no preferred orientation and wide variations in size are found between individuals of the same species. The corals present in these claystones also are randomly oriented, suggesting biogenic mixing. Coral epithecae are frequently corroded and encrusted by multiple generations of epibionts (auloporids, bryozoans, worm tubes, and inarticulate brachiopods). Typically this corrosion and encrustation is confined to one side of the coral, indicating long term exposure at the sediment-water interface with infrequent reorientation. Some corals and brachiopods are reduced to very small, abraded roll fragments suggesting several episodes of agitation and transportation. The fossil material in the claystones is poorly sorted and mud-supported. No primary sedimentary structures are present.

Preservation in the limestones is considerably better. Most brachiopods remain articulated; proetid trilobites are complete and typically found with their cephala and thoracopygidia at oblique angles to one another (a presumed semi-infaunal life orientation); corals are uncorroded, with many found in an upright life orientation; whole crinoid calyces with attached stems have been found in these limestones, as well as recrystallized mollusk shells that usually exhibit moldic preservation. Fossils exhibit little or no compaction and are less closely packed than those in synjacent claystones. All these taphonomic features support a quick-burial, smothered-bottom interpretation for these limestones (see below).

Overall faunal evidence suggests deposition in shallow, slightly agitated water in euphotic environments, within the reach of storm wave base. The evidence of biocorrosion (by algae or other organisms) and encrustation of corals suggests very low rates of sedimentation. Times of episodic rapid sedimentation also must have existed. Fistuliporoid bryozoan colonies showing signs of partial death due to sediment smothering and then regeneration, have been collected from these samples.

The excellent preservation in the two lower limestone beds (samples 11, 13) of the *Heliophyllum — Heterophrentis* fauna indicates a rapid burial by storm-suspended mud for these beds. The sudden blanketing of benthic communities by mud layers may have resulted in locally raised pH or altered sediment chemistry which, in turn, favored precipitation of early diagenetic carbonate cements (Waage, 1964; Speyer and Brett, 1983).

Association 5: *Favosites hamiltoniae*

The fauna of this association is very similar to that of the large rugose coral association with one major addition, the large tabulate coral *Favosites hamiltoniae*. Although *F. hamiltoniae* is rarely found in the large rugose coral association, it is common in this association and the colonies found here are much larger (the largest colony found has a diameter of 50 cm). Large solitary rugosans, the colonial rugose coral *Eridophyllum*, and ramose favositids are all present to common. However, the small rugosans, *Stereolasma* and *Amplexiphyllum*, are absent. Most of the brachiopods present also are found in the large rugose coral association, the most common being *Mediospirifer, Elita, Meristella, Rhipidomella,* and *Longispina*. There are a few diagnostic additions including *Fimbrispirifer*, *Spinocyrtia* and *Centronella*. Much crinoid and fenestrate bryozoan debris is present.

Diversity levels in this association are moderate, being slightly lower than those calculated for the large rugose coral association (see Fig. 7 and Table 1 for sample 19). Relative abundances are low and equitable for most species (Figs. 8 and 9).

This association is found in the central argillaceous limestone (sample, 19 Fig. 7). This grain-supported calcisiltite is intensely bioturbated and has no sedimentary structures. The association corresponds with part of biofacies 'C2' of the binary-data clustering results (Fig. 7).

Fossils here are typically disarticulated (articulation ratios much less than 1) and highly fragmented and abraded. Crinoid and bryozoan fragments make up much of the debris. Brachiopods frequently show roll faceting. The corals present are usually biocorroded and encrusted by epibionts. Here, however, in contrast to the large rugose coral association, most corals have their epitheca uniformly corroded and encrusted. The massive favositid heads commonly are overturned. Fossils have a random orientation.

All the evidence points to a moderate- to high-energy, agitated environment, near normal wave base. Robust corals like *F. hamiltoniae* are believed to have thrived in a shallow, shoal setting. Partial winnowing of fine-grained sediments together with reorientation of large fossils indicates slightly greater turbulence than was characteristic of the large-rugose-coral biofacies, but conditions were otherwise quite similar.

DISCUSSION

Lithologic and faunal cyclicity is quite apparent for the Centerfield interval. Moreover, the nearly perfect symmetry is of foremost importance in accounting for this cyclicity. With the sole exception of the two lower "event limestone beds" (Salt Creek and Browns Creek beds), the faunal and lithologic changes are perfectly mirrored about the middle limestone (Schaeffer Creek beds). The middle limestone, therefore, represents a reversal point in faunal and lithologic change, and, by implication, probably an endpoint of an environmental spectrum.

Depositional Models

Three alternative depositional models have previously been suggested to explain similar faunal and lithologic sequences found elsewhere in the Hamilton Group. Any of the three could be invoked to explain the Centerfield cycle, but only one of these models appears consistent with stratal data. The first model considers fluctuations in environmental parameters independent of depth as a cause for facies shifts, while the other two call for changes in water depth, sedimentation rate, or both.

The first explanation was specifically developed for the dark-gray and black shales and mudstones of the Ledyard Member of the Ludlowville Formation which possess many of the same faunal associations as does the Centerfield interval (McCollum, 1980). McCollum envisions a restricted epeiric-sea for Ludlowville time, wherein a great deal of freshwater runoff and high organic productivity led to a stratified water column. Depth throughout the basin was fairly uniform, never exceeding a shallow sublittoral limit. According to this model, changes in faunal assemblages throughout the Ludlowville shales do not necessarily reflect a gross change in water depth but rather changes in oxygen, substrate, and salinity as controlled by the input of fresh water and sediment. Carbonate deposition is explained by periods of low-clastic sedimentation.

The second alternative has been developed independently several times for various Hamilton shale and carbonate packages. In this interpretation, emphasis has been placed on the origin of the limestone "key beds," including the Centerfield limestone, which mark the bases of the four formations within the Hamilton Group. This model calls for a rapid rise in sea level (a eustatic transgression) causing alluviation or coastal entrapment of sediments. The corresponding reduction of clastic sedimentation seen in the basin's interior would lead to the deposition of dyscyclic deep water carbonates. The bases of these limestones, therefore, would represent isochronous, disconformable surfaces that cross-cut facies (Cooper, 1957). This explanation has been applied to the Tully Limestone (Friedman and Johnson, 1968), the Portland Point complex (McCave, 1968, 1969a, 1969b, 1973), the Mottville Limestone (Wolff and Buttner, 1980), and to some extent, the Centerfield Limestone (Kramers, 1970; McCave, 1973).

Although this "transgression-alluviation model" for Hamilton limestone beds does not directly deal with the depositional history of the dark shales, a shallow-water origin is implied (Cooper, 1957). A shallow-water interpretation for black shales has been applied to the Upper Devonian, Chattanooga Shale (Conant and Swanson, 1961) and Pennsylvanian shales associated with marine cyclothems (Zanger and Richardson, 1963; Merrill and Martin, 1976; von Bitter, 1972).

The third alternative, and the one advocated in this study of the Centerfield limestone, invokes a shallow-water origin for many of the Hamilton limestone beds and a deeper water origin for the dark-gray and black shales. In contrast to the transgression-alluviation model, this model postulates that the limestones were deposited near or at times of peak regression, when fine terrigenous clastics were being winnowed from the shallow-water area and being transported to the deeper basinal settings. This "regression-winnowing hypothesis" for shallow-water limestone and the deeper basinal dark shales has been applied to the Middle Devonian Mahantango Formation (upper Hamilton equivalent) in Pennsylvania (Ellison, 1965; Faill and others, 1978), to the Upper Devonian rocks in New York (Sutton and others, 1970; Thayer, 1974; Rickard, 1981), and even to Pennsylvanian cyclothems (Heckel, 1977; Malinki, 1982).

Several recent studies of the Hamilton Group in New York (Brett and Baird, 1985) have suggested that the major key limestone beds were deposited in shallow-water. Most of these studies have relied on regional stratigraphic evidence; thus shale-to-carbonate transition in western New York has been correlated with an obvious coarsening-upward shale to sandstone hemicycle in eastern New York (Selleck and Hall, 1977; Selleck, 1983; Baird and Brett, 1983a). Other types of paleontologic (Boardman, 1960) and sedimentologic (Muskatt, 1964) studies have suggested similar interpretations for many of these shale-limestone transitions. Only one study has specifically dealt with the depositional history of the Centerfield. Kramers (1970) has suggested that the Centerfield limestone was deposited as a result of sediment trapping behind an eastern shoal and suggests this may have been accompanied by a minor regression, but does not present any evidence for this regressive interpretation.

Many lines of evidence derived from the present study suggest that the regressive-winnowing hypothesis holds true for the Centerfield interval, and these lines also counter the other two alternatives as possible explanations. This paper focuses mainly on faunal arguments, but other taphonomic, sedimentologic, and stratigraphic arguments also support this interpretation, as discussed below.

Faunal Evidence

All of the faunal patterns seen throughout the Centerfield interval are consistent with a shallow water-carbonate and deeper-water, black-shale interpretation. Patterns of diversity, relative abundances, and biofacies distributions can all be explained with this reconstruction of depositional history. Very few modern-day bodies of water can be used as models for stratified epeiric-sea environments, but possible analogs include the Black Sea and California borderland basins (Savrda and others, 1984). Studies in these Recent land-locked waters with restricted circulation, yet normal-marine salinities, have shown a low-high-low pattern in faunal diversity. Low diversities are expected in very shallow-water, stressed areas within normal wavebase (< 20 m). Shallow subtidal areas, below normal wavebase in well aerated waters, should have higher diversities with an abundant shelly fauna. In stratified basins, the deeper subtidal settings become dysaerobic to anaerobic (or oxygen limiting) and, although other environmental parameters may well be stable and predictable, faunal diversities decline correspondingly. The progressive depletion of oxygen in bottom waters and sediments toward the stratified basin center should lead to a stress gradient which increases basinward. Oxygen levels between about 0.3 and 0.5 ml/l are too low to support an abundant shelly infauna, yet high enough to support both a shelly epifauna and infaunal soft-bodied taxa (Boucot, 1981; Thompson and others, 1985).

This model seems to fit the diversity pattern seen throughout the Centerfield Member. The lowest diversities are seen within the dark, fissile shales that have been interpreted as being the deepest-water setting (see below). The *Leiorhynchus* — styliolinid biofacies associated with this lithology is dominated by a low-diversity, thin shelled benthic brachiopod and pelagic fauna. The type of fauna present and the diversity level, therefore, further support this environmental interpretation. The calcareous claystones, on the other hand, exhibit a high species richness, dominated by both a rich shelly epifauna and infauna. Coral-rich biofacies are interpreted as representing shallow subtidal environments. The Centerfield Member, in this geographic area, does not show any evidence for reaching the shallowest, within-normal-wavebase environments. Therefore, the other end of the low-diversity, high-stress spectrum of the Rhoads and Morse model cannot be seen. Such conditions are seen, however, in other associations with lower diversities in the temporally equivalent Chenango Sandstone of central New York State. The roughly cyclic pattern in diversity (see Fig. 7) can, therefore, be interpreted as a shift in depositional environments resulting from shallowing and deepening in a stratified basin.

The appearance of opportunistic and specialist species at their respective positions along the section is also predicted by the environmental reconstructions given above. A dominance of opportunistic species is expected in stressed environments such as those hypothesized for the *Leiorhynchus* — styliolinid and the *Ambocoelia* — chonetid associations (Levinton, 1970; Alexander, 1977; Hickey and Younker, 1981). Low stress offshore environments should show high diversities of stenotopic taxa, as is the case for the large rugose coral and *Favosites hamiltoniae* associations. The *Pseudoatrypa* — *Stereolasma* association exhibits intermediate diversity and equitability and a mixture of opportunistic and specialist species, indicative of conditions intermediate between high-stress (low oxygen) deeper-water and stable, shallower shelf settings.

The faunal associations and their corresponding inferred paleoecological settings found within the Centerfield (discussed earlier) are also consistent with the environmental reconstructions for similar faunal associations seen elsewhere in other Silurian and Devonian rocks. Ambocoeliid-dominated communities are typically found in facies indicative of deeper epeiric sea settings (Benthic Assemblage 4 or 5; Boucot, 1975), whereas coral-rich favositid communities are common in shallow subtidal areas (Benthic Assemblage 2). Atrypid-rich communities are found at intermediate depths (Benthic Assemblage 3). Large rugose and favositid coral-dominated assemblages are known to occur elsewhere within the Hamilton Group and other studies dealing with these coral-beds have reached a similar conclusion concerning environment (Oliver, 1951; Beerbower, 1960; Grasso, 1968, 1973; Baird and Brett, 1983b). It is also interesting to note that many of the same coral genera seen in these Hamilton coral beds are also found in Onondaga (Edgecliff Member, lower Middle Devonian) bioherms, many of which show evidence of upward growth into settings of high current activity (Crowley and Poore, 1974; Feldman, 1980; Koch, 1981; Lindemann and Feldman, 1981; Oliver, 1960; Williams, 1980).

Taphonomic and Sedimentologic Evidence

Taphonomic and sedimentologic evidence (outlined in the previous section) also supports the regressive carbonate-transgressive, dark-shale interpretation. There is circumstantial evidence that virtually all Centerfield sediments were deposited in waters of the photic zone. However, there are also sedimentologic indications of variation in environmental energy, and hence depth, throughout the cycle.

Muskatt's (1964) petrologic study of the Hamilton Group in the Syracuse area (approximately mid-way between the type sections for both the Centerfield and Chenango Members) demonstrates an increase in detrital-quartz content and in mean grain size (0.02 to 0.045 mm) towards the center of the Centerfield cycle. The predominantly silt-sized matrix of the middle limestone in the Genesee Valley area suggests persistent current activity at least strong enough to winnow out most of the clay-sized sediment. Evidence for deposition at or near normal wavebase exists in the Centerfield middle limestone of Genesee County; this unit is comparable to the Tichenor Limestone (Buchanan and others, 1972; Baird, 1979). Thus, taphonomic and sedimentologic evidence suggests a shallow-water depositional environment, between normal- and storm-wave bases, for the middle limestone (<25 m).

Dark-gray shale endpoints of the Centerfield cycle, on the other hand, possess taphonomic and sedimentologic features implying the opposite — a deeper water origin (>100 m). Finely laminated clayshales lacking other primary sedimentary structures, coupled with an absence of bioturbation, suggests very low-energy levels. Well preserved delicate fossils, low degrees of fragmentation, and an absence of obvious storm layers are consistent with a deep, quiet-water setting. The upper and lower claystone transitions show an ordered gradation of intensity for all these features and, therefore, represent intermediate depths.

It is notable that the probable storm-generated beds occur only in sediments of biofacies C and not in biofacies A and B (Fig. 7). Thus, evidence for deposition at or near storm wavebase is indicated only near the center of the cyclic package. Again, this may indicate shallower water conditions than are found in the black shales.

Stratigraphic Evidence

In addition to the sedimentologic and paleontologic evidence, stratigraphic evidence strongly supports a shallow-water interpretation of the middle argillaceous limestones of the Centerfield Member. Detailed mapping of Centerfield equivalents east of the Genesee Valley has revealed that the lower shale to middle limestone gradation of western New York correlates with an upward-coarsening hemicycle in eastern New York (Gray, 1984), much like the hemicycles previously discussed for other Hamilton beds (Muskatt, 1964). The Centerfield middle limestone (Schaeffer Creek Submember) is thus a direct correlative of the crossbedded Chenango Sandstone, while the underlying Levanna Shale correlates with the shales and fine siltstones of the Butternut Member. Correlation with a shallowing (coarsening-upward) clastic interval in central New York implies a similar shallowing in the western part of the state for the lower shale-claystone-limestone Centerfield transition.

Stratigraphic work to the west has shown one interesting change in the Centerfield middle limestone where the outcrop belt fortuitously moves northward in Genesee County, further up the inferred paleoslope. The central limestone at White Creek is represented by a coarse grained encrinite, indicating an even shallower environment, perhaps within normal wavebase (Gray, 1984).

Implications

In the context of this shallow-water limestone and deep water black-shale interpretation, the entire Centerfield sequence represents a regressive-transgressive cycle reflected in the changes seen in fauna and lithology. The Centerfield middle limestone represents the center of symmetry where environmental trends begin to reverse themselves. The lower regressive hemicycle begins with a quiet offshore dysaerobic setting below storm wavebase (100 – 150 m) and culminates with the deposition of the limestone core in shallow, within storm wavebase depths (20 – 25 m). It is then followed by the transgressive hemicycle and a return to dysaerobic deeper water.

Cyclicity seen for the faunal associations and the quantitatively defined biofacies pattern suggest that benthic communities migrated in response to the changing water depth. Assuming the existence of an east-west depositional strike in western New York (see Introduction), whole faunal associations would have migrated southward (downslope), during the lower regressive hemicycle, tracking their optimum habitat type. Conversely, during the transgressive phase of the cycle, faunal associations were tracking northward (upslope) as the water deepened. The large rugose coral and *Favosites hamiltoniae* associations, for example, would have been found slightly further north at the time the *Pseudoatrypa* – *Stereolasma* association was thriving in the area represented by the present day Genesee Valley outcrop belt. During times when the *Ambocoelia* – chonetid and *Leiorhynchus* – styliolinid associations were dwelling in this area, the coral-dominated associations probably would have existed still farther north of the present outcrop belt.

The five faunal associations found throughout the Centerfield interval are not confined to this level but recur throughout the Hamilton Group at various levels, especially within the Ludlowville and Moscow Formations of western and central New York (Miller, 1982; Kloc, 1983; Grasso, 1973, this volume; Brett, 1974a, 1974b; Baird and Brett, 1983b). Preliminary cluster analyses done for many samples of the Hamilton Group in western and central New York show a consistency in biofacies for Ludlowville and Moscow time.

Not only do these associations recur at various levels within the western New York Hamilton Group, but they also show a sequential ordering of appearance with respect to each other, similar to that observed in Centerfield interval. Thus, they are also components of similar faunal and lithologic cycles (Brett and Baird, 1985). These other Hamilton cycles may or may not be centered about an argillaceous limestone or have the same perfect symmetry as seen for the Centerfield. However, in many cases, coral-dominated associations appear near the centers of the cycles and similar ordered progressions of associations can be seen on either side of the centers. Cycles showing an asymmetry (associations or lithologies missing from the ordered sequence) are commonly interrupted by obvious discontinuity surfaces.

The existence of several comparable cycles within the Hamilton Group suggests that similar regressive-transgressive episodes occurred throughout Hamilton depositional history, and that ancient communities merely tracked these changes. The consistency in the faunal association constituents also suggests an overall "community stasis" for much of Hamilton history in this part of New York. The cause of these small scale regressive-transgressive cycles is unclear. However, the widespread occurrence of these rhythms and their apparent synchroneity, at least throughout the Appalachian Basin, suggests these changes may be eustatic in nature (House, 1983).

ACKNOWLEDGMENTS

A number of people associated with the University of Rochester were involved with this project at various stages and we would like to thank all of them. Many students, most notably Fritz Hertel, helped with the many painstaking hours of bulk sampling. Dr. Judy Massare helped with the clustering program and offered many helpful suggestions on the early drafts of the manuscript. Frank Ettensohn and Peter von Bitter critically reviewed this paper and offered many helpful suggestions. Finally, the field-related portion of this project was made possible by grants from the Rochester Museum and Science Center, NSF grant EAR 8313103 and from the donors to the Petroleum Research Fund of the American Chemical Society.

REFERENCES CITED

Alexander, R. 1977. Growth, morphology and ecology of Paleozoic and Mesozoic opportunistic species of brachiopods from Idaho-Utah. J. Paleontol. 51(6):1133-1149.

Baird, G.C. 1979. Sedimentary relationships of Portland Point and associated Middle Devonian rocks in central and western New York. New York State Mus. Bull. 433, 23 p.

_________ 1981. Submarine erosion of a gentle paleoslope: a study of two discontinuities in the New York Devonian. Lethaia 14:105-122.

_________ and Brett, C.E. 1981. Submarine discontinuities and sedimentary condensation in the upper Hamilton Group (Middle Devonian): examination of marine shelf and paleoslope deposits in the Cayuga Valley. *In* New York State Geol. Assn. Guidebook, 53rd Ann. Mtg., SUNY at Binghamton, p. 115-145.

_________ and _________ 1983a. Middle Devonian detrital shelf-to-basin ramp: Ludlowville lithofacies and faunal associations in west-central New York. Geol. Soc. Amer. Abstr. with Programs 15(3):170.

_________ and _________ 1983b. Regional variation and paleontology of two coral beds in the Middle Devonian Hamilton Group of western New York. J. Paleontol. 57(3):417-446.

Beerbower, J.R. 1960. Paleoecology of the Centerfield coral zone, East Stroudsburg locality, Monroe County, Pennsylvania. Penn. Acad. Sci. Proc. 34:84-91.

_________, Buehler, E.J., Bray, R.G., and Jordan, F.W. 1969. A "paleomicroecologic" study of Mid-Devonian marine assemblages (Hamilton Group) in western New York. North Amer. Paleontol. Conven., Field Trip Guidebook 5:48-78.

_________ and Jordan, D. 1969. Application of information theory to paleontologic problems: taxonomic diversity. J. Paleontol. 43(5):1184-1198.

Belak, R. 1980. The Cobleskill and Akron Members of the Rondout Formation: Late Silurian carbonate shelf sedimentation in the Appalachian Basin, New York. J. Sediment. Petrol. 50:1187-1204.

Boardman, R.S. 1960. Trepostomatous bryozoa of the Hamilton Group of New York State. U.S. Geol. Survey Prof. Paper 340, 87 p.

Boucot, A.J. 1975. Evolution and Extinction Rate Controls. Developments in Palaeontology and Stratigraphy 1. Elsevier Press, Amsterdam, 427 p.

_________ 1981. Principles of Benthic Marine Paleoecology. Academic Press, New York, 463 p.

Brett, C.E. 1974a. Biostratigraphy and paleoecology of the Windom Shale Member (Moscow Formation) in Erie County, New York. *In* New York State Geol. Assn. Guidebook, 46th Ann. Mtg., SUNY College at Fredonia, p. G1-15.

_________ 1974b. Contracts of the Windom Member (Moscow Formation) in Erie County, New York. *In* New York State Geol. Assn. Guidebook, 46th Ann. Mtg., SUNY College at Fredonia, p. C1-29.

_________ 1983. Sedimentology, facies and depositional environments of the Rochester Shale (Silurian; Wenlockian) in western New York and Ontario. J. Sediment. Petrol. 53:947-971.

_________ and Baird, G.C. 1982. Upper Moscow-Genesee stratigraphic relationships in western New York: evidence for regional erosive beveling in the late Middle Devonian. *In* New York State Geol. Assn. Guidebook, 54th Ann. Mtg., SUNY at Buffalo, p. 19-63.

_________ and Baird, G.C. 1985. Carbonate-shale cycles in the Middle Devonian of New York: An evaluation of models for the origin of limestones in terrigenous shelf sequences. Geology 13:324-327.

_________, Baird, G.C., and Kloc, G.J. 1983. Paleoenvironmental models of the Hamilton Group: cyclic and episodic deposition in a stratified basin. Geol. Soc. Amer. Abstr. with Programs 15(3):172.

_________ and others. 1983. Middle Devonian (Givetian) coral associations of western New York State. *In* Sorauf, J.E. and Oliver, W.A., Jr., eds., Silurian and Devonian Corals and Stromatoporoids of New York, 4th Intl. Symp. on Fossil Cnidaria, p. 65-107.

Buchanan, H., Couch, E.L., and Gibbons, V.A. 1972. Paleoenvironment of the Tichenor Limestone of the Ludlowville Formation of western New York. Geol. Soc. Amer. Abstr. with Programs 4:4-5.

Cheetham, A.H. and Hazel, J.E. 1969. Binary (presence-absence) similarity coefficients. J. Paleontol. 43:1130-1136.

Cleland, H.F. 1903. A study of the Hamilton Formation of the Cayuga Lake section in central New York. U.S. Geol. Surv. Bull. 206, 112 p.

Clifford, H.T. and Stephenson, W. 1975. An Introduction to Numerical Classification. Academic Press, New York, 229 p.

Conant, L.C. and Swanson, V.E. 1961. Chattanooga Shales and related rocks of central Tennessee and nearby areas. U.S. Geol. Surv. Prof. Paper 357, 91 p.

Cooper, G.A. 1930. Stratigraphy of the Hamilton Group of New York. Amer. J. Sci. 19:116-134, 214-236.

_________ 1957. Paleoecology of the Middle Devonian of eastern and central United States. *In* Ladd, H.S., ed., Treatise on Marine Ecology and Paleoecology, 2, Paleoecology. Geol. Soc. Amer. Mem. 67:249-278.

_________ and others. 1942. Correlation of the Devonian sedimentary formations of North America. Geol. Soc. Amer. Bull. 53:1729-1794.

Crowley, D.J. and Poore, R.Z. 1974. Lockport (Middle Silurian) and Onondaga (Middle Devonian) patch reefs in western New York. *In* Peterson, D.N., ed., New York State Geol. Assn. Guidebook, 46th Ann. Mtg., SUNY College at Fredonia, p. A-1 to A-41.

Davis, J.C. 1973. Statistics and Data Analysis in Geology. John Wiley and Sons, New York, 550 p.

Dennison, J.M. and Hasson, K.O. 1976. Stratigraphic cross section of Hamilton Group (Devonian) and adjacent strata along south border of Pennsylvania. Amer. Assn. Petroleum Geol. Bull. 60(2):278-287.

_________ and Head, J.W. 1975. Sea level variations interpreted from the Appalachian Basin Silurian and Devonian. Amer. J. Sci. 275(10):1089-1120.

Ellison, E.L. 1965. Stratigraphy and paleontology of the Mahantango Formation. Penn. Topog. Geol. Surv., Gen. Geol. Rep. G-48, 298 p.

Ettensohn, F.R. and Barron, L.S. 1981. Depositional model for the Devonian-Mississippian black shales of North America: a paleoclimatic-paleogeographic approach. *In* Roberts, T.G., ed., Geol. Soc. Amer. Guidebook, 1981 Field Trip, Cincinnati; Amer. Geol. Inst., 2:344-361.

_________ 1982. A tectonic-climatic approach to the deposition of the Devonian-Mississippian black-shale sequence of North America. Proc. 1982 Eastern Oil Shale Symposium. Kentucky Dept. Energy and Univ. Kentucky Inst. for Mining and Minerals Research, Lexington, Kentucky, p. 5-37.

Faill, R.T., Hoskins, R.T., and Wells, R.B. 1978. Middle Devonian stratigraphy in central Pennsylvania — a revision. Penn. Geol. Surv., Gen. Geol. Rep. 70, 20 p.

Farris, J.S. 1969. On the cophenetic correlation coefficient. Systematic Zoology 18:279-285.

Feldman, H. R. 1980. Level-bottom brachiopod communities in the Middle Devonian (Onondaga) of New York. Lethaia 13:27-46.

Fernow, L.R. 1961. Paleoecology of the Middle Devonian Hamilton Group in the Cayuga Lake region. Doctoral dissertation, Cornell University, 209 p.

Flessa, K.W. and Bray, R.G. 1977. On the measurement of size-independent morphological variability: an example using successive populations of a Devonian spiriferid brachiopod. Paleobiology 3:350-359.

Fox, W.T. 1968. Quantitative paleoecologic analysis of fossil communities in the Richmond Group. J. Geol. 76:613-640.

Friedman, G.M. and Johnson, K.G. 1968. The Devonian Catskill deltaic complex of New York, type example of a tectonic delta complex. *In* Shirley, M.L. and Ragdale, J.A., eds., Deltas in their Geologic Framework, Houston Geol. Soc., Houston, Texas, p. 189-211.

Grabau, A.W. 1898-1899. Geology and paleontology of Eighteen Mile Creek and the lake shore sections of Erie County, New York. Buffalo Soc. Nat. Sci. Bull. 6, 403 p.

_________ 1899. The faunas of the Hamilton Group of Eighteen Mile Creek and vicinity in western New York. New York State Mus. Ann. Rep. 16:227-239.

Grasso, T.X. 1968. A new coral bed in the Hamilton Group (Middle Devonian) of central New York. J. Paleontol. 42(1):84-87.

_________ 1970. Paleontology, stratigraphy, and paleoecology of the Ludlowville and Moscow Formations (Upper Hamilton Group) in central New York. *In* New York State Geol. Assn. Guidebook, 42nd Ann. Mtg., SUNY College at Cortland, p. D11-22.

_________ 1973. A comparison of environments: the Middle Devonian Hamilton Group in the Genesee Valley: faunal analysis, Jaycox Run. *In* New York State Geol. Assn. Guidebook, 45th Ann. Mtg., SUNY College at Brockport, p. B1-24.

_________ 1983. Stratigraphy and depositional environments of the Mottville Member in central and eastern New York. Geol. Soc. Amer. Abstr. with Programs 15(3):170.

_________ (this volume). Redefinition, stratigraphy, and depositional environments of the Mottville Member (Hamilton Group) in central and eastern New York.

Gray, L.M. 1983. Biofacies and depositional environments of the Centerfield "Member" in the northern Appalachian basin. Geol. Soc. Amer. Abstr. with Programs 15(3):170.

_________ 1984. Lithofacies, biofacies, and depositional history of the Centerfield Member (Middle Devonian) of western and central New York State. Doctoral dissertation, University of Rochester, 158 p.

Harland, W.B., Cox, A.V., Llewellyn, P.G., Pickton, C.A.G., Smith, A.G., and Walters, R. 1982. A geologic time scale. Cambridge Earth Science Series, Cambridge Univ. Press, 131 p.

Heckel, P.H. 1977. Origin of black shale facies in Pennsylvanian cyclothems of mid-continent North America. Amer. Assn. Petroleum Geol. Bull. 61:1045-1068.

Hickey, D.R. and Younker, J.L. 1981. Structure and composition of a Pennsylvanian benthic community. J. Paleontol. 55(1):1-12.

House, M.R. 1983. Devonian eustatic events. Proc. Ussher Soc. 5:396-405.

Kauffman, E.G. 1977. Evolutionary rates and biostratigraphy. *In* Kauffman, E.G. and Hazel, J.E., eds., Concepts and Methods of Biostratigraphy. Dowden Hutchinson and Ross, Stroudsburg, Pa., p. 109-142.

Kissling, D.L. and Mosshier, S.O. 1981. Subsurface Onondaga Limestone: Stratigraphy, facies and paleogeography. *In* Enos, P., ed., New York State Geol. Assn. Guidebook, 53rd Ann. Mtg., SUNY at Binghamton, p. 279-280.

Kloc, G. 1983. Stratigraphic distribution of ammonoids from the Middle Devonian Ludlowville Formation in New York. Master's thesis, SUNY at Buffalo, 73 p.

Koch, W.F. 1978. Brachiopod paleoecology, paleobiogeography, and biostratigraphy in the late Middle Devonian of eastern North America. Doctoral dissertation, Oregon State Univ., 122 p.

_________ 1981. Brachiopod community paleoecology, paleobiogeography, and depositional topography of the Devonian Onondaga Limestone and correlative strata in eastern North America. Lethaia 14:83-103.

Kramers, W.J. 1970. The Centerfield limestone (Middle Devonian) of New York State and its clastic correlatives: a sedimentological analysis. Doctoral dissertation, Rensselaer Polytech. Inst., 126 p.

Laub, R.S. 1972. The auloporid genus *Cladochonus* McCoy, 1847: new data from the New York Devonian. J. Paleontol. 46(3):364-370.

Levinton, J.S. 1970. The paleoecologic significance of opportunistic species. Lethaia 3:69-78.

Lindemann, R.H. and Feldman, H.R. 1981. Paleocommunities of the Onondaga Limestone (Middle Devonian) in central New York State. *In* New York State Geol. Assn. Guidebook, 53rd Ann. Mtg., SUNY at Binghamton, p. 79-93.

Lindholm, R.C. 1969. Carbonate petrology of the Onondaga Limestone of New York: a case for calcisiltite. J. Sediment. Petrol. 39:268-275.

Lloyd, M. and Ghelardi, R.J. 1964. A table for calculating the "equitability" component of species diversity. Amer. Nat. 98:190-192.

McCave, I.N. 1968. Shallow and marginal marine sediments associated with the Catskill complex in the Middle Devonian of New York. Geol. Soc. Amer. Spec. Paper 106, p. 75-107.

_________ 1969a. Correlation of marine and non-marine strata with an example from the Devonian of New York. Amer. Assn. Petroleum Geol. Bull. 53:155-162.

_________ 1969b. Correlation using a sedimentological model of part of the Hamilton Group (Middle Devonian) of New York. Amer. J. Sci. 267:567-591.

_________ 1973. The sedimentology of a transgression: Portland Point and Cooksburg Members (Middle Devonian), New York State. J. Sediment. Petrol. 43(2):484-504.

McCollum, L.B. 1980. Distribution of marine fossil assemblages in a Middle Devonian stratified basin, lower Ludlowville Formation, New York. Doctoral dissertation, SUNY at Binghamton, 143 p.

McGhee, G.R. and Sutton, R.G. 1981. Late Devonian marine ecology and zoogeography of the central Appalachians and New York. Lethaia 14:27-43.

MacDonald, K.B. 1975. Quantitative community analysis: recurrent group and cluster techniques applied to the fauna of the Upper Devonian Sonyea Group, New York. J. Geol. 82:473-499.

Malinki, J. 1982. Paleontology and depositional environments of Pennsylvanian low oxygen shales. Mid continent North America. Geol. Soc. Amer. Abstr. with Programs 14(7):554.

Mello, J.F. and Buzas, M.A. 1968. An application of cluster analysis as a method of determining biofacies. J. Paleontol. 42(3):747-758.

Merrill, G.K. and Martin, M.D. 1976. Environmental control of conodont distribution in the Bond and Mattoon Formations (Pennsylvanian, Missourian), Northern Illinois. Geol. Assn. Canada Spec. Paper 15:243-271.

_________ and von Bitter, P.H. 1976. Revision of conodont biofacies nomenclature and interpretation of environmental controls in Pennsylvanian rocks of eastern and central North America. Royal Ontario Museum Life Sciences Contrib. 108, 46p.

Miller, K.B. 1982. Paleomicroecologic analysis of depositional environments and sequences, "*Pleurodictyum* zone," Ludlowville Formation, Middle Devonian of western New York. Master's thesis, SUNY at Binghamton, 131 p.

_________ (this volume). Depositional environments and sequences, "*Pleurodictyum* zone," Ludlowville Formation of western New York.

Muskatt, H.S. 1964. Petrology of the siltstones in the Ludlowville Formation (Middle Devonian), Onondaga County, New York. *In* New York State Geol. Assn. Guidebook, 36th Ann. Mtg., Syracuse University, p. 116-124.

Oliver, W.A., Jr. 1951. Middle Devonian coral beds of central New York. Amer. J. Sci. 249(10):705-728.

_________ 1960. Coral faunas in the Onondaga Limestone of New York. U.S. Geol. Surv. Prof. Paper 400-B, p. 172-174.

Pielou, E.C. 1966. Shannon's formula as a measure of species diversity: its use and misuse. Amer. Nat. 100(914):463-465.

Rhoads, D.C. and Morse, J.W. 1971. Evolutionary and ecologic significance of oxygen-deficient marine basins. Lethaia 4(4):413-428.

Rickard, L.V. 1975. Correlation of the Silurian and Devonian rocks of New York State. New York State Mus. Map and Chart Ser. 24, 16 p.

_________ 1981. The Devonian System of New York State. *In* Oliver, W.A., Jr. and Klapper, G., eds., Devonian Biostratigraphy of New York. Intl. Union Geol. Sci., Subcom. on Devonian Strat. Part 1., Washington, D. C., p. 5-22.

Savarese, M. 1984. Faunal and lithologic cyclicity in the Centerfield Member (Middle Devonian: Hamilton Group) of western New York: a reinterpretation of depositional history. Master's thesis, University of Rochester, 110 p.

Savrda, C.E., Bottjer, D.J. and Donn, S.G. 1984. Development of a comprehensive euxinic biofacies model: evidence from Santa Monica, San Pedro and Santa Barbara Basins, California continental borderland. Amer. Assn. Petroleum Geol. Bull. 68:1179-1192.

Selleck, B. 1983. Cyclic shallowing-upward sequences in the Hamilton Group of central New York. Geol. Soc. Amer. Abstr. with Programs 15(3):170.

Selleck, B. and Hall, R. 1977. Sedimentology and paleontology of portions of the Hamilton Group in central New York. *In* New York State Geol. Assn. Guidebook, 49th Ann. Mtg., SUNY at Oneonta, p. 1-23.

Shannon, C.E. and Weaver, W. 1963. The Mathematical Theory of Communication. University of Illinois Press, Urbana, Illinois, 117 p.

Smith, B. 1935. Geology and mineral resources of the Skaneateles Quadrangle. New York State Mus. Bull. 300, 120 p.

Sneath, P.H.A. and Sokal, R.R. 1973. Numerical Taxonomy. W.H. Freeman and Co., San Francisco, 573 p.

Speyer, S.E. and Brett, C.E. 1983. Storm-generated "event horizons" in the Middle Devonian Hamilton Group of western New York. Geol. Soc. Amer. Abstr. with Programs 15(3):171.

Stephenson, W., Williams, W.T., and Cook, S. 1972. Computer analyses of Petersen's original data on bottom communities. Ecol. Monogr. 42:387-415.

Sutton, R.G., Bowen, Z.P., and McAlester, A.L. 1970. Marine shelf environments of the Upper Devonian Sonyea Group of New York. Geol. Soc. Amer. Bull. 81:2975-2992.

Thayer, C.W. 1974. Marine paleoecology in the Upper Devonian of New York. Lethaia 7:121-155.

__________ 1981. Ecology of living brachiopods. *In* Broadhead, T.W., ed., Lophophorates: Notes for a Short Course. University of Tennessee, Department of Geological Sciences, Studies in Geology 5, p. 110-126.

Thompson, J.B., Mullins, H.T., Newton, C.R., and Vercoutere, T.L. 1985. Alternative biofacies model for dysaerobic communities. Lethaia 18:167-179.

Valentine, J.W. 1973. Evolutionary Paleoecology of the Marine Biosphere. Prentice-Hall, Englewood Cliffs, New Jersey, 511 p.

Valentine, J.W. and Peddicord, R.G. 1967. Evaluation of fossil assemblages by cluster-analysis. J. Paleontol. 41:502-507.

von Bitter, P.H. 1972. Environmental control of conodont distribution in a Shawnee Group (Upper Pennsylvanian) of eastern Kansas. Univ. Kansas Paleontol. Contrib. Art. 59, 105 p.

Waage, K.M. 1964. Origin of repeated fossiliferous concretion layers in the Fox Hills Formation. *In* Merriam, D.F., ed., Symposium on Cyclic Sedimentation, Kansas Geol. Surv. Bull. 169:541-563.

Williams, L.A. 1980. Community succession in a Devonian patch reef (Onondaga Formation, New York) — physical and biotic controls. J. Sediment. Petrol. 50:1169-1185.

Wolff, M.P. and Buttner, P.J.R. 1980. Marine and fluvial delta platform environments of the transgressive clastic correlatives of the Middle Devonian (Erian) Mottville Limestone Member of the Skaneateles Formation in eastern New York State. *In* New York State Geol. Assn. Guidebook, 52nd Ann. Mtg., Rutgers University, p. 243-271.

Zangerl, R. and Richardson, E.S., Jr. 1963. The paleontological history of two Pennsylvanian black shales. Fieldiana, Geol. Mem. 4, 352 p.

Depositional Environments and Sequences, "*Pleurodictyum* Zone," Ludlowville Formation of Western New York

Keith B. Miller[1]
Department of Geological and Environmental Sciences
State University of New York
at Binghamton

ABSTRACT

Seven faunal assemblages, defined by cluster-analysis using presence-absence taxonomic data, occur within a three meter interval of highly fossiliferous shales and thin limestones at the base of the Wanakah Member, Ludlowville Formation, of the New York Middle Devonian. These assemblages occur in a vertically symmetrical sequence which records a regressive-transgressive event in which the assemblages represent faunal zones paralleling a carbonate shelf to the north. Taphonomic and sedimentological evidence indicates the increasing effects of storm activity toward the middle of the section representing the time of maximum regression. Storm deposits were critical to the development of the rich epifaunal suspension-feeding assemblages characteristic of the central part of the section. Following each rapid colonization of the storm layers, burrowing activity of infaunal deposit-feeders combined with pulsational mud deposition to progressively exclude the epifauna. Despite the rich epifauna, a soft substrate and reducing conditions within the sediment are indicated for nearly the entire studied interval. The results of this investigation suggest that detailed paleoecological studies of mud rocks have great potential for refining, and in some cases altering, existing environmental reconstructions.

INTRODUCTION

This paper reports a detailed study of the paleoecology and depositional environment of a three-meter interval of fossiliferous gray calcareous shales and thin argillaceous limestones within the Hamilton Group, Middle Devonian (Erian) of western New York. Nearly a century of stratigraphic and paleontological work has revealed considerable lithologic and faunal variation in the Hamilton, and has recognized several distinctive, widespread beds. The "*Pleurodictyum* Beds" and "Trilobite Beds" of Grabau (1899a), comprising the "*Michelinia* Zone" of Cleland (1903), are examples of such beds and the focus of this study. These beds define the base of the Wanakah Shale member of the Ludlowville Formation near the middle of the Hamilton shale sequence. Their diverse fossil assemblages are anomalies within the thick sequence of Alden, Elma, and Wanakah shales of western New York (McCollum, 1981) and occur again only in association with the prominent widespread Tichenor and Centerfield Limestones. Through detailed understanding of the *Pleurodictyum* and Trilobite Beds, additional light can be shed on the factors influencing the observed faunal and lithologic character of the Middle Devonian shales and limestones of New York.

This investigation isolates some of the biotic, abiotic, and taphonomic factors controlling the composition of muddy level bottom marine fossil assemblages. A detailed study of the taxonomy, taphonomy, and sedimentology of mud rocks permits the retrieval of valuable information about the depositional environment which is overlooked by studies done on a coarser scale. In the absence of sedimentary structures, the taphonomy and autecology of fossil organisms can reveal much about the physical environment. In addition, a more detailed understanding of the nature and extent of lateral and vertical variability enables a more realistic reconstruction of ecological and depositional processes to be made. To this end, the fauna and enclosing shale were studied at a scale near the limit of stratigraphic resolution, here defined as the sediment thickness within which no regular vertical variability can be recognized (i.e., sedimentation units). The beds defined in the field typically contained several sedimentation units each with distinctive fossil accumulations, or taphocoenoses, differing in taphonomic as well as taxonomic characteristics. The stratigraphic descriptions were supplemented by sampling adjacent, well exposed, bedding plane surfaces on a square decimeter scale. Because of slow depositional rates and bioturbation, however, even these surfaces represent a "paleontologic time exposure" (Schindel, 1980) and not an instant in time. Geologically very short time intervals could be studied nevertheless, and temporal changes in the physical or biological environment were recorded on a scale perhaps not much greater than that which would occur during a human lifetime.

---

[1] Present address, Department of Geological Sciences, The University of Rochester.

GENERAL DEPOSITIONAL SETTING

The general paleogeographic and depositional setting of the Hamilton Group is reasonably well understood. Confined to an area east and southeast of the Findlay-Algonquin Arch, the Hamilton black and gray shales and thin limestones were deposited in a semi-enclosed lobe of an extensive Middle Devonian tropical and subtropical epeiric sea which covered most of eastern North America between an emergent Transcontinental Arch and the rising Appalachians (Oliver, 1976). The eastern and southeastern margins of this lobe were dominated by the prograding clastic shorelines of the Catskill delta complex. The sea's northwestern shore was probably a gently sloping crinoidal carbonate platform which now is represented only by the Tichenor and Portland Point Members (McCave, 1973; Baird and Brett, 1981b).

An actively subsiding northeast-southwest trending trough appears to have been centered in the Canandaigua and Seneca Valleys during the deposition of the Wanakah Member and underlying Ledyard Member of the Ludlowville Formation. The stratigraphic relationships of this interval have been recently defined and revised by Baird (1979). In the upper Wanakah equivalent of the King Ferry Member, Baird and Brett (Baird and Brett, 1981b; Baird, 1981) have proposed the existence of a northwest gently-dipping paleoslope associated with the southeast side of this trough in the Cayuga Valley. Reflecting this paleoslope, the "*Pleurodictyum* Zone" becomes increasingly detrital as it approaches the clastic dominated eastern basin margin, and is represented on the southeastern shore of Cayuga Lake by a thick coarsening upward clastic interval with scattered *Pleurodictyum* culminating in cross-bedded fine to medium grained sandstone (Brett and others, 1983). On the opposite side of the basin axis, the Wanakah and Ledyard Members both exhibit a facies transition from gray to black shale from Lake Erie toward the Canandaigua and Seneca Valleys. The *Pleurodictyum* and Trilobite Beds, although displaying a progressive eastward decrease in faunal diversity from Erie County toward the presumed basin center (McCollum, 1981), remain gray and fossiliferous. This suggests that the rate of subsidence of the trough, and its absolute depth, varied significantly with time. The axis of deposition was also not constant with time, but migrated westward to the Genesee Valley during the subsequent deposition of the Deep Run and Kashong Shale Members. The westward and eastward thinning of these beds and the Tichenor is a consequence of sedimentary condensation, non-deposition, and submarine erosion. Total absence of the Jaycox, Deep Run, and Kashong Shale Members from Erie County may suggest the existence of a carbonate shoal there at that time (Baird and Brett, 1981b). While there is no evidence of emergence in Erie County during Ledyard and Wanakah deposition, the Hamilton sea clearly shallowed as it approached the Lake Erie region.

The depositional interval of the Ludlowville and Moscow formations is characterized by several small regressive-transgressive cycles within the overall regional regressive trend resulting from the progradation of the Catskill clastic wedge. The Tichenor-Portland Point, Centerfield, and Menteth Limestones are interpreted by Baird and Brett (1981a) to record such regressive episodes. This explanation differs from McCave's (1973) transgressive model for these deposits, but seems to be clearly supported by their detailed sedimentologic work and the results of this study.

METHODS AND PRINCIPLES

Field Studies

The three meter section of the Wanakah Shale Member investigated includes the "*Strophalosia*," "*Nautilus*," "*Pleurodictyum*," and "Trilobite Beds" of Grabau (1899a). As defined by Cooper (1930) the Wanakah begins at the base of the "*Strophalosia* Bed," which is an easily recognizable thin limestone unit. The other diagnostic beds of the studied interval are surprisingly persistent along the strike of the outcrop belt. The "Trilobite Beds," which lie at the top of the interval, can be traced from Lake Erie eastward to Canandaigua, and the "*Pleurodictyum* Zone" continues as far as Cayuga Lake (Cooper, 1930).

Two outcrops in Erie County, New York were selected for study. They are both located between the towns of Lancaster and East Aurora about 16 km southeast of Buffalo in the Depew 15′ quadrangle. The primary study area was along Buffalo Creek 1 km north of the Bullis Road bridge (Sites 1 and 2 on Fig. 1). At Site 1, five stratigraphic sections were located along a 120 meter segment of stream bank exposures. A sixth section was measured downstream at Site 2 where a high cut bank exposes the entire studied interval for a few hundred meters. The second area lay on Cazenovia Creek 8.6 km to the west along the outcrop belt and about 0.8 km downstream from the Transit Road bridge (Site 3 on Fig. 1). Here one section was measured along a 40 meter exposure, and a second at a cut bank about 30 meters downstream.

In order to retrieve the greatest amount of sedimentologic and paleontologic information, it was necessary to obtain the best stratigraphic resolution possible. The true characterization of vertical changes further required that taxonomic and taphonomic data be collected in the field over several lateral meters of outcrop exposure. Vertical sequences of bulk samples, while yielding a wealth of paleontological data in the laboratory (Ferguson, 1962), would not take into account lateral variability. If fossil distributions were patchy, the additional taxonomic information obtained by laboratory examination could not be unambiguously interpreted in terms of temporal change. Therefore, stratigraphic sections were described in the field on the basis of the detailed examination of approximately

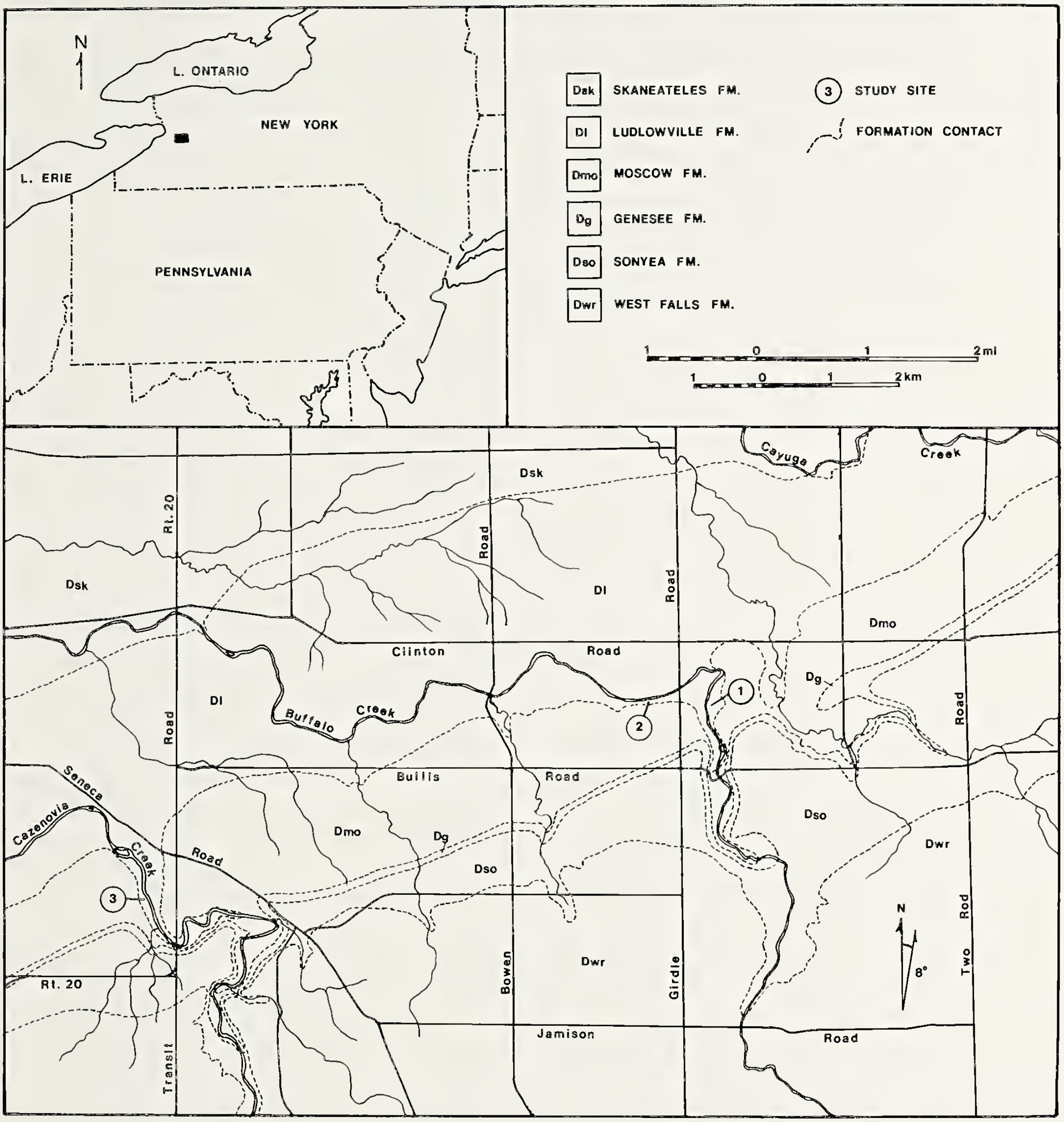


Figure 1 Devonian outcrop pattern in the East Aurora and Elma area of Erie County, and location of study sites (geology after Buehler & Tesmer, 1963).

five-meter-wide portions of vertical outcrop. By pooling fossil data over this distance, small scale patchiness was not as easily confused with vertical change. Information on small scale lateral variability was obtained through the description and sampling of available bedding plane surfaces. Larger scale patchiness was recognized by comparing nearby vertical sections.

For each of the stratigraphic sections, beds were defined by lithology (primarily fissility) and faunal composition. The thinnest intervals which could be traced over at least five meters of outcrop were recorded. These beds ranged from two to three centimeter highly fossiliferous layers to intervals of over 30 centimeters in poorly fossiliferous and homogeneous shales.

All the macrofossil species observed within the beds over approximately five lateral meters were recorded. The reliability of the presence-absence data obtained, however, was dependent on the duration of the examination. For this reason, the sections were reexamined several times to compensate for the improvement in field identification with time. This problem has been dealt with similarly by other workers (Palmer and Palmer, 1977). Taphonomic information on orientation, articulation, and fragmentation also was compiled for each bed. Clustering, if recognized, was described, aided by the examination of adjacent bedding plane exposures when available. The beds thus characterized seldom equaled the limits of stratigraphic resolution of the section, and their contained fossils, therefore, represented assemblages, not taphocoenoses. Collection of large blocks of the thin limestones within the sections, however, permitted the sedimentation units and taphocoenoses of those beds to be studied through acetate peels.

Because of the difficulty of defining "community" limits and the inconsistent use of the term in the literature (Stanton, 1976), an alternative operational terminology is used here. A taphocoenosis, as understood in this paper, is the total group of fossils occurring together within the same sedimentation unit. If post-mortem transportation is not significant and if condensation and bioturbation is minimal, a taphocoenosis would represent a biocoenosis or life assemblage minus the unpreserved elements. Adapting Kauffman's (1973) usage, assemblages are defined as accumulations of fossils from one or more biocoenoses or taphocoenoses, and associations as specific recurrent groups of fossils which were components of biocoenoses. Associations may persist through several different successive biocoenoses. The ability to distinguish between assemblages, biocoenoses, and associations is important to the accurate paleoecologic interpretation of fossil accumulations.

In addition to the presence-absence and taphonomic data collected for the vertical sections, several bedding plane surfaces correlated with beds within these measured sections were sampled. Laboratory analysis of the samples involved disaggregation of the shale and the counting and measurement of all identifiable skeletal material. As well as giving some insight into the extent of lateral variability within beds, the data thus obtained gave a more accurate picture of faunal composition, density, and relative abundance than was possible from examination of the outcrop alone. Also, the size frequency data provided valuable information about the population dynamics of individual species which in turn revealed additional information about the environment. A thorough discussion of the results of these analyses can be found elsewhere (Miller, 1982).

Faunal Analyses

The "*Pleurodictyum* Zone" has a very abundant and diverse fauna in contrast to the relatively fossil-poor shales immediately above and below that are dominated by *Devonochonetes* and *Mucrospirifer*. Well over sixty species of macrofossils can be identified, and they include rugose and auloporid corals, bryozoans, pelmatozoans, bivalves, gastropods, and a diverse assemblage of brachiopods. Such a rich fauna is not encountered again in the Wanakah until the "*Demissa*" and "*Stictopora*" Beds just below the Tichenor Limestone (Grabau, 1899a).

The assemblages recorded are believed to be essentially *in situ* as suggested by the taphonomy of their component species. The only fossil benthos believed to have been significantly transported are broken and abraded crinoid ossicles. Orthocone cephalopods, and probably styliolinids, are pelagic and, therefore, cannot be tied directly to the benthic assemblages in which they occur.

The combined effects of non-preservation, dissolution, and the non-recognition of small or rare species are potentially great (Johnson, 1964; Lawrence, 1968), and impose limitations on the interpretation of ancient ecosystems. They do not, however, prohibit the recognition of assemblages with distinct ecologic and environmental significance (Ziegler, 1973). Assemblages representing intervals of stratigraphically contiguous sedimentation units containing similar taphocoenoses can be easily identified. Inspection of the presence-absence data suggests that the three meter section can be subdivided into several such distinct intervals, but in order to define their boundaries on a more objective and repeatable basis, cluster analysis was used.

Similarity values between pairs of samples and pairs of species were calculated using the Jaccard Coefficient. This coefficient, which ignores common absences, is a widely used similarity measure for presence-absence data (MacDonald, 1975; Mello and Buzas, 1968; Cheetham and Hazel, 1969). Clusters were generated with a FORTRAN IV weighted-pair group averaging program and printed with a dendrogram subroutine from Davis (1973). A published program for generating binary similarity matrices written by Millendorf and others (1978) was also utilized as a subroutine.

For a composite Buffalo Creek Section, a Q-mode analysis produced a computer generated dendrogram with a cophenetic correlation coefficient (Kaesler, 1970) of 0.819. The complete presence-absence faunal data for this stratigraphic section are present in Figure 2. Only those species occurring in more than one sample were placed in the data matrix used in the cluster analysis. Inspection of the resultant dendrogram, shown in Figure 3, reveals that the section can be subdivided into six assemblages occurring through stratigraphically contiguous beds. The sequence and stratigraphic ranges of these assemblages, which are here named by two or more of their characteristic taxa, are illustrated in Figure 4. The *Styliolina*-Ostracod Assemblage comprises the shale and two thin limestone beds above the lowest "Trilobite Bed," which, together with the shale bed underlying it, belongs to the *Aulocystis-Stereolasma* Assemblage. The most highly fossiliferous portion of the section, extending down to and including

	BCVS1-1	BCVS1-2	BCVS1-3	BCVS1-4a	BCVS1-4b	BCVS1-5	BCVS1-6	BCVS1-7	BCVS1-8	BCVS1-9	BCVS1-10	BCVS3-11	BCVS3-12	BCVS3-13	BCVS3-14	BCVS3-15	BCVS3-16	BCVS3-17	BCVS3-18	BCVS3-19	BCVS6-6	BCVS6-7	BCVS6-8	BCVS6-9	BCVS6-10	BCVS6-11	BCVS6-12	BCVS6-13	BCVS8-14
Ambocoelia nana																												X	X
A. umbonata	X		X	X	X	X	X	X	X	X						X													
Athyris spiriferoides					X			X	X	X	X	X	X	X	X				X		X								
Camarotoechia sappho														X				X											
Craniella hamiltoniae						X				X																			
Cryptonella rectirostra					X		X		X	X	X		X	X															
Cyrtina hamiltonensis										X				X															
Devonochonetes coronatus	X		X	X	X	X	X	X	X	X	X	X	X	X		X	X	X	X	X	X			X	X		X	X	X
Douvillina inequistriata											X		X																
Elita fimbriata								X																					
Mediospirifer audaculus								X	X	X	X	X	X	X	X			X											
Megastrophia concava			X					X	X	X	X	X	X									X							
Mucrospirifer mucronatus	X	X	X	X	X	X	X	X	X	X	X	X	X	X	X	X		X	X		X	X	X	X			X	X	
Orbiculoidea lodiensis				X	X					X																			
Pholidostrophia nacrea					X				X																				
Protoleptostrophia perplana			X	X		X		X	X	X	X			X								X							
Rhipidomella penelope			X			X	X	X	X		X	X	X																
R. cyclas ?					X			X	X	X	X																		
Schuchertella arctostriatus																													
Spinocyrtia granulosa					X				X		X			X															
Spinulicosta sp.									X	X																			
Tropidoleptus carinatus				X																								X	
Cypricardella bellistriata			X						X	X																			
Cypricardinia indenta					X				X	X	X	X																	
Modiomorpha sp.						X						X																	
Nucula corbuliformis																													
Paleoneilo sp.				X	X				X	X				X				X											X
Pterinopecten sp.																													
Ptychopteria sp.				X	X									X															
Pseudaviculopecten princeps																													
Naticonema lineata						X	X	X		X		X	X	X															
Paleozygopleura delphicola																												X	X
Platyceras sp.														X															
Retispira leda											X																		X
Pleurotomaria lucina																													
Styliolina fissurella	X	X	X																									X	X
Tentaculites bellulus																													
Spyroceras nuntium														X														X	X
Dipleura dekayi													X																
Greenops boothi	X		X	X	X									X															
Phacops rana	X	X	X	X	X	X	X	X	X	X	X	X	X	X	X			X		X		X						X	
OSTRACODS	X	X	X	X					X	X																			
Aulocystis commensalis									X					X															
A. cooperi														X															
A. dichotoma				X	X	X	X	X	X	X	X					X		X	X	X	X								
A. jacksoni					X	X				X		X	X																
Aulopora microbuccinata								X	X				X								X								
Amplexiphyllum hamiltoniae													X																
Cystiphylloides sp.													X																
Favosites nitella ?													X																
Heliophyllum halli													X																
Pleurodictyum americanum														X		X		X		X									
P. dividua ?																													
Stereolasma rectum					X	X	X	X	X	X	X	X	X					X		X									
Trachypora romingeri														X															
Hederella sp.						X						X																	
FENESTRATE BRYOZOANS					X	X	X	X	X	X	X	X	X	X	X			X	X	X									
RAMOSE BRYOZOANS					X			X	X	X	X	X	X	X	X	X		X	X	X	X	X	X						
MASSIVE & ENCRUSTING BRYOZOANS										X		X	X								X	X							
Ancyrocrinus bulbosus								X																					
CRINOIDS					X	X		X	X	X	X	X	X	X	X			X			X								

Figure 2 Presence-absence faunal data for composite Buffalo Creek section used to generate cluster analysis dendrogram.

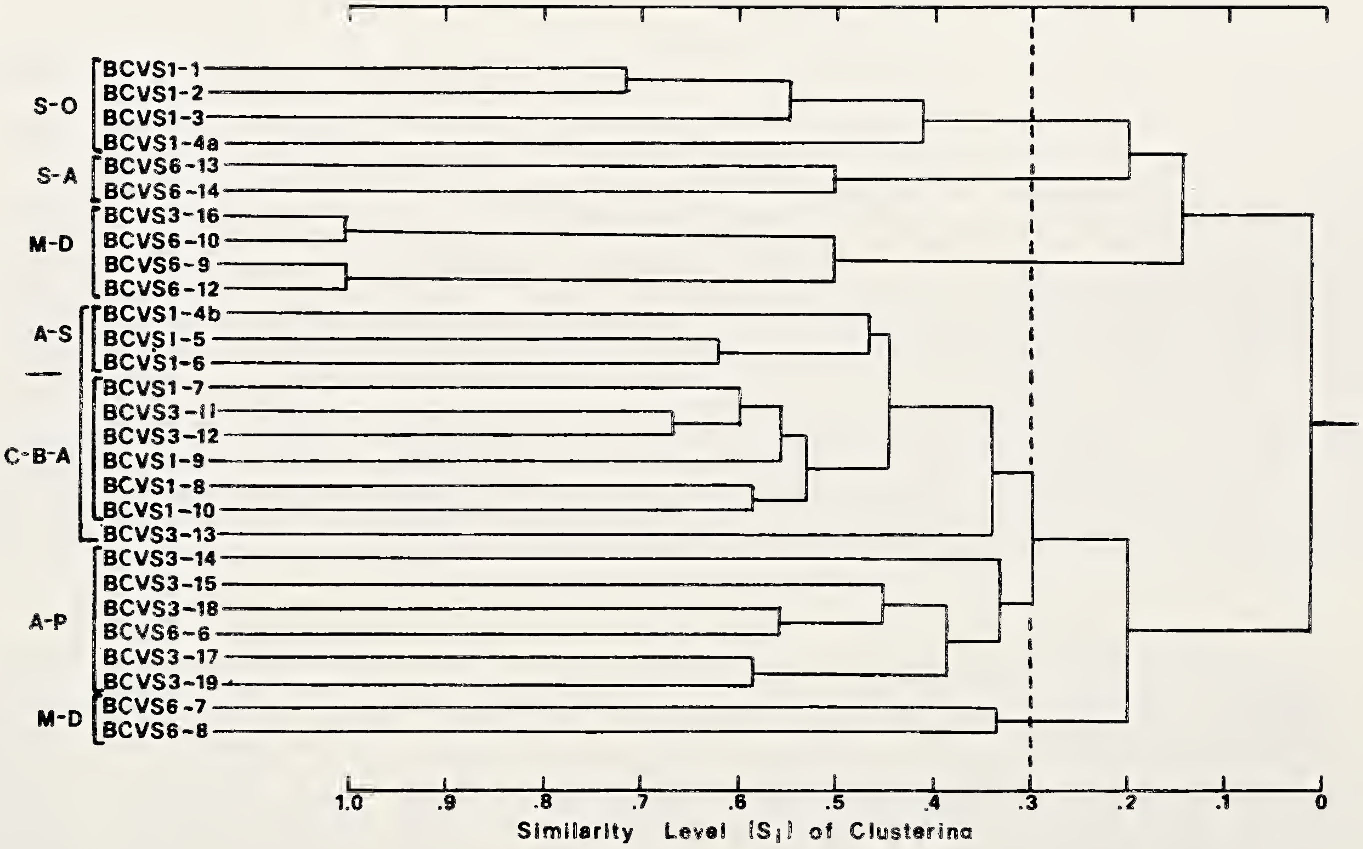


Figure 3 Computer generated Q-mode dendrogram for composite Buffalo Creek section using Jaccard similarity coefficients (Sj). S-O) *Styliolina*-Ostracod Assemblage, S-A) *Styliolina-Ambocoelia* Assemblage, M-D) *Mucrospirifer-Devonochonetes* Assemblage, A-S) *Aulocystis-Stereolasma* Assemblage, C-B-A) Crinoid-Bryozoan-*Athyris* Assemblage, A-P) *Aulocystis- Pleurodictyum* Assemblage.

a pseudo-concretionary layer, is assigned to the Crinoid-Bryozoan-*Athyris* Assemblage. The underlying *Pleurodictyum*-containing shales are grouped into the *Aulocystis-Pleurodictyum* Assemblage. Below this the *Mucrospirifer-Devonochonetes* Assemblage comprises the interval of fossil-poor shales. Grabau's "*Strophalosia* Bed" together with the overlying *Ambocoelia*-rich concretionary layer belong to the *Styliolina-Ambocoelia* Assemblage. The same general pattern of faunal units just described was generated by a Q-mode analysis of the Cazenovia Creek sections.

The beds named and described by Grabau (1899a) are easily recognizable markers which enable the studied interval to be readily subdivided in the field. The assemblages defined above place these marker beds within the context of a coherent pattern of faunal change. Previous descriptive work failed to disclose this pattern, the most significant aspect of which is the similarity of the assemblages above and below the Crinoid-Bryozoan-*Athyris* Assemblage. Table 1 summarizes the characteristic faunal content for each of the assemblages.

TAPHONOMY AND ENVIRONMENTAL INTERPRETATION OF ASSEMBLAGES

The three meter section studied possesses a considerable degree of faunal symmetry with the *Styliolina*, *Mucrospirifer-Devonochonetes*, and *Aulocystis* Assemblages flanking the central Crinoid-Bryozoan-*Athyris* Assemblage above and below. I believe that the taphonomic and sedimentologic evidence strongly suggest that these assemblages are depth related and represent faunal belts paralleling an ancient shoreline. Their vertical pattern is, therefore, interpreted as a relatively rapid regressive-transgressive event.

Styliolina-Ambocoelia Assemblage

Lithology and fossil content — This assemblage is characteristic of the basal beds of the studied interval beginning with an argillaceous lime mudstone, correlative with Grabau's "*Strophalosia* Bed," in which the small cricoconarid *Styliolina* is abundant. The gastropods *Palaeozygopleura* and *Retispira* and orthoconic cephalopods are common. Diffuse-bounded, vertical burrows about 2 – 3 mm or greater in diameter are present. Overlying this bed is a bioclastic wackestone to packstone dominated by the small spiriferid brachiopod *Ambocoelia nana* and by *Styliolina*. Branching concretionary bodies about 9 cm wide characterize this bed and are especially well displayed at Buffalo Creek. The brachiopods common to both of these beds (i.e. *Ambocoelia*, *Tropidoleptus*, and *Devonochonetes*) are of unusually small size, except for the somewhat less common *Mucrospirifer*.

Taphonomy — The brachiopods of this assemblage are excellently preserved with a large number of articulated specimens, some of which are geopetally filled. Within the branching *Ambocoelia*-rich concretions the densely-packed

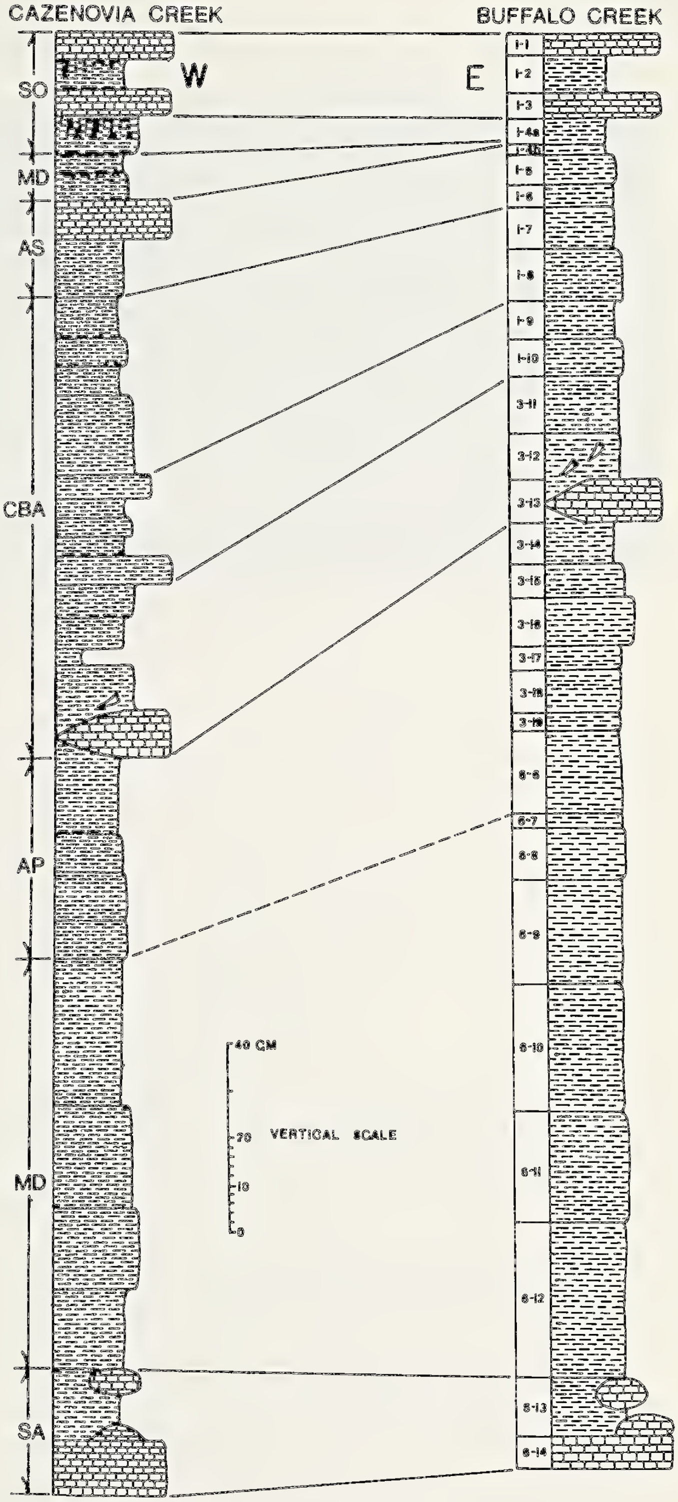


Figure 4 Stratigraphic sections for composite Buffalo and Cazenovia Creek sections. Bed numbers for Buffalo Creek section from figure 2. SO) Styliolina-Ostracod Assemblage, AS) *Aulocystis-Stereolasma* Assemblage, CBA) Crinoid-Bryozoan-Athyris Assemblage, AP) *Aulocystis-Pleurodictyum* Assemblage, MD) *Mucrospirifer-Devonochonetes* Assemblage, SA) *Styliolina-Ambocoelia* Assemblage.

Table 1
Faunal composition of assemblages

Assemblages	
Styliolina-Ostracod	*Styliolina fissurella*
	Phacops rana
	ostracods
	Mucrospirifer mucronatus
	Devonochonetes scitulus
Mucrospirifer-Devonochonetes	*Mucrospirifer mucronatus*
	Devonochonetes scitulus
Aulocystis-Stereolasma	*Aulocystis dichotoma*
	Stereolasma rectum
	Phacops rana
	Ambocoelia umbonata
	Pholidostrophia nacrea
	Naticonema lineata
	Ancyrocrinus bulbosus
Crinoid-Bryozoan-*Athyris*	crinoids
	bryozoans
	Stereolasma rectum
	Athyris spiriferoides
	Rhipidomella penelope
	Mediospirifer audaculus
	Spinocyrtia granulosa
	Megastrophia concava
	Protoleptostrophia perplana
"coral bed"	*Heliophyllum halli*
	Trachypora romingeri
	fistuliporid bryozoan mounds
Aulocystis-Pleurodictyum	*Aulocystis dichtoma*
	Pleurodictyum americanum
	Mucrospirifer mucronatus
	Devonochonetes scitulus
Mucrospirifer-Devonochonetes	*Mucrospirifer mucronatus*
	Devonochonetes scitulus
Styliolina-Ambocoelia	*Styliolina fissurella*
	Ambocoelia nana
	Mucrospirifer mucronatus
	Devonochonetes scitulus
	Tropiodoleptus carinatus
	Paleozygopleura delphicola
	Spyroceras nuntium

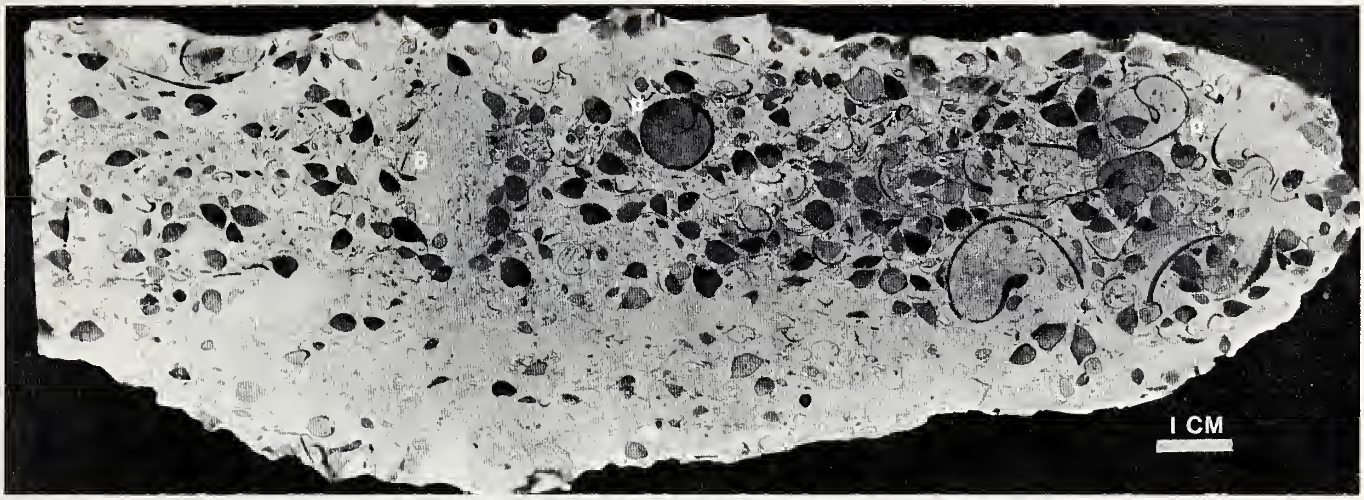


Figure 5 **Photograph of a longitudinal section of concretionary body from acetate peel. Several paleozygopleurid gastropods (p) visible in right half of photograph and a diffuse bounded vertical burrow (B) can be seen at left.**

articulated shells are randomly oriented and show no evidence of compaction. Brachiopod shell hash, together with styliolinids, is abundant within the matrix. Orthoconic cephalopods are typically inclined with their apices pointed downward.

In the "*Strophalosia* Bed" fine grained pyrite occurs within gastropod and orthoconic cephalopod shells and as irregular masses probably associated with burrows. Pyrite is also present in the concretionary *Ambocoelia* horizon, disseminated in the matrix and concentrated in the micritic fillings of some articulated shells. Most articulated shells in both beds are filled with calcite spar.

Environmental interpretation — The very small size of the common brachiopods of this assemblage, together with the high dominance and low diversity of the fauna, suggests unfavorable conditions for all but a few benthic shelled taxa. Early maturation and rapid reproduction associated with small opportunistic forms, would be advantageous in an environment with fluctuating oxygen levels. McCollum (1981) interprets her "diminutive faunas," which include the *Styliolina-Ambocoelia* Assemblage of this study, as representing poorly oxygenated conditions, and applies the stratified oxygen deficient basin model of Rhoads and Morse (1971) to the Ludlowville sea. Similarly, Bowen and others (1974) have postulated a zone of low and fluctuating oxygen below the "mixed and fully aerated surface waters" of the Upper Devonian Sonyea sea. Together with Thayer (1974) they report the common occurrence of *Ambocoelia* in the prodelta and open shelf muds lying within this low oxygen zone.

The excellent preservation of the external detail of brachiopod shells and the abundance of articulated specimens clearly indicates an *in situ* fauna. This conclusion is supported by the lack of sorting of the fossil material. Relatively quiet water conditions, therefore, are indicated. Some winnowing of mud by very weak bottom currents may have contributed to the concentration of shell material in the *Ambocoelia* concretionary horizon (Fig.5). A period of very low sediment influx, however, was probably responsible for the formation of this bed. This is suggested by the high concentration of the tiny cricoconarid *Styliolina*, whose pelagic habit is indicated by its ubiquitous occurrence in all Hamilton facies and its abundance in otherwise fossil-poor shales.

Though the foregoing evidence suggests relatively deep water, the occurrence of the herbivorous gastropods *Palaeozygopleura* and *Retispira* indicates the presence of algae and, therefore, a depth within the photic zone. According to Peel (1978), high spired loxonemataceans, like *Palaeozygopleura*, were likely microherbivores on algal coated surfaces and "probably moved freely in soft sediments." He argues against any gastropod with maximum size over 20 mm being epiplanktonic, and most specimens of *Palaeozygopleura* from this assemblage are several centimeters long. *Retispira* may also have been a microherbivore, and foliage dwelling cannot be ruled out.

The petrology of the limestones and concretions indicates both a fluid substrate and reducing conditions within the sediment. The existence of a fluid substrate is supported by the diffuse boundaries of burrows which indicate thixotropic flow rather than plastic deformation of the sediment during burrowing (Rhoads, 1970). The apex down orientation of orthoconic cephalopods is also consistent with this interpretation. Disseminated pyrite and pyritic shell fillings point to reducing conditions within the sediment, which favored the accumulation of organic matter and the growth of sulfate reducing bacteria. Anaerobic bacterial reduction of sulfate, using organic matter as an energy source, is the principle source of sulfide for pyrite formation (Berner, 1970). Berner has proposed that elemental sulfur is required for the conversion of FeS to FeS_2. His suggestion that S is generated by the oxidation of H_2S in the sediment by the diffusion or mixing of oxygen from overlying aerobic waters implies an oxidation/reduction boundary at, or just below, the sediment-water interface. The formation of concretions was also, like pyrite, probably a response to the anaerobic decomposition of organic matter. The branching *Ambocoelia* concretions were clearly formed very early in diagenesis as demonstrated by the absence of shell distortion from compaction as seen in Figure 5. The generation of ammonia during decomposition results in both an increase of pH and the formation of carbonate and bicarbonate ions (Berner, 1968; Jordan, 1968).

Both the faunal composition and taphonomy of this assemblage are very similar to that of the pyritic fossil beds of the Hamilton Group as described by Dick (1982). The low energy, dysaerobic environments with "soupy" bioturbated substrates inferred by Dick for the pyritic beds, agree substantially with the interpretation made here for the *Styliolina-Ambocoelia* Assemblage. This assemblage may, therefore, represent "marginally favorable" conditions at the transition between anaerobic and oxygenated substrates in a stratified basin.

Lower *Mucrospirifer-Devonochonetes* Assemblage

Lithology and fossil content — Overlying the low diversity fossil-rich basal beds, is an interval of poorly fossiliferous gray shale containing widely scattered *Mucrospirifer* and *Devonochonetes* shells. Rare bryozoans and crinoid ossicles can be found. There are, however, several thin horizons which contain brachiopods, such as *Athyris*, *Rhipidomella*, *Megastrophia*, and *Protoleptostrophia*, and corals, such as *Aulocystis* and *Stereolasma*, characteristic of beds higher in the section. These horizons are best displayed at the Cazenovia Creek outcrop. *Zoophycos* spreiten as well as some vertical burrows are present, and pyritized burrows sometimes form the nuclei of small carbonate concentrations. Large subcircular concretions also are present locally within this shale interval.

Taphonomy — The episodic coral and brachiopod bearing horizons are developed on thin hashy pavements of shell material of species occurring in the intervening shales. *Devonochonetes* is dominant, but *Mucrospirifer*, *Phacops*, and *Styliolina* also are common components. Stereolasmid corals rest directly on the shell pavements and some occur in life position. Partially articulated crinoid stems are common. These horizons frequently are somewhat pyritized and often contain pyrite nodules; a few form thin hard calcareous bands.

Environmental interpretation — The lower *Mucrospirifer-Devonochonetes* Assemblage probably represents a depositional environment below wave base quite similar to that of the *Styliolina-Ambocoelia* Assemblage. The generally very low faunal density and diversity of this interval was largely a consequence of the absence of suitable attachment sites. *Mucrospirifer* and *Devonochonetes* were well adapted for the prevailing soft bottom conditions. The alate "wings" of the former and the concavoconvex shape and hinge spines of the latter helped prevent foundering on soft substrates.

The importance of available attachment sites is indicated by the restriction of the corals and larger brachiopods of this interval to a few thin shell hash pavements. These shell pavements, being composed primarily of ubiquitous brachiopods and pelagic styliolinids, likely represent winnowed horizons produced by unusually severe storms that were able to affect the bottom well below wave base. The patchy attachment surfaces produced by winnowing were then available for colonization by a brachiopod and coral fauna that was otherwise excluded. The brachiopod clusters of the upper Wanakah studied by Bray (1971) were similarly related to isolated attachment surfaces. The occurrence of *Athyris*, *Rhipidomella*, strophomenids, and small corals suggest that oxygen levels were somewhat higher than for the *Styliolina-Ambocoelia* Assemblage, though the sediment remained anoxic. The pyritization and occasional carbonate enhancement of the fossiliferous horizons suggests rapid burial by a mud layer, probably of storm origin, resulting in the decomposition of organic matter under anoxic conditions within the sediment.

Aulocystis-Pleurodictyum Assemblage

Lithology and fossil content — The shale above the *Mucrospirifer-Devonochonetes* Assemblage has several very fossiliferous horizons containing the small discoid tabulate coral *Pleurodictyum*. This interval is correlative with the upper two *Pleurodictyum* Beds of Grabau (1899b), and ranges from very fissile gray calcareous shales to soft nodular shales. The fossiliferous layers are dominated by and sometimes nearly composed of the tiny tabulate coral *Aulocystis*. The intervening shales are fossil-poor and characterized by *Mucrospirifer* and *Devononchonetes*.

Taphonomy — The beds of this shale interval show considerable faunal variability over lateral distances of only a few tens of meters. Fossil-rich layers vary in thickness as well as faunal composition from section to section, an abundance of *Aulocystis* being their only persistent characteristic. Large scale faunal patchiness would seem to be suggested. On a smaller scale, within the fossil-rich layers, *Mucrospirifer* occurs in small clusters about 10 cm in diameter, and *Aulocystis* occurs in well defined clusters.

Thin skeletal hash layers of *Devonochonetes*, *Ambocoelia*, and crinoid ossicles form pavements underlying the fossil-rich layers in a manner similar to that of the *Mucrospirifer-Devonochonetes* Assemblage. With the notable exception of *Athyris*, brachiopods tend to be disarticulated and convex up. About 50 percent of the *Pleurodictyum* coralla were found in an inverted position, and some also show marked constrictions followed by renewed growth.

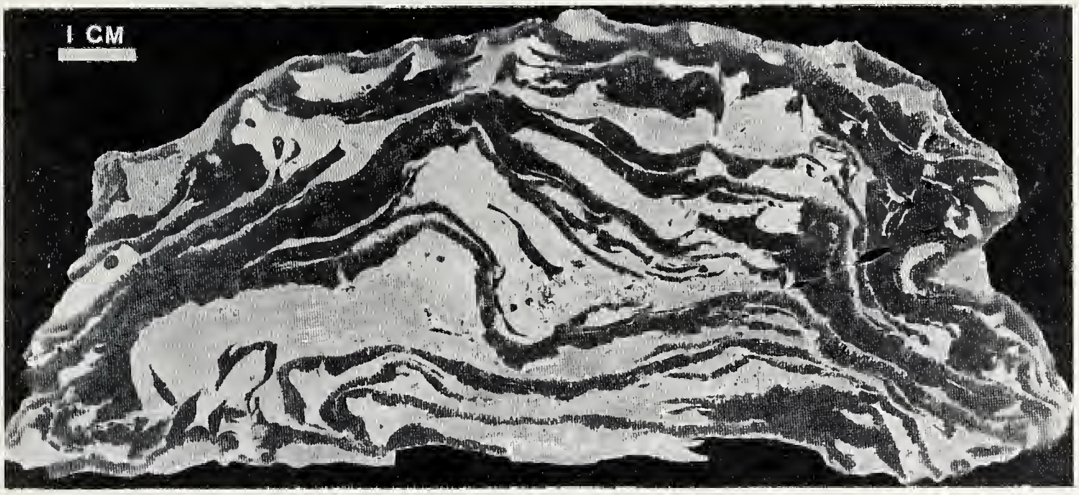


Figure 6 Photograph of a large hemispherical fistuliporid bryozoan mound from Buffalo Creek made from an acetate peel showing internal sediment layers.

Environmental interpretation — Both the clustering and larger scale patchiness are strong evidence of an *in situ* and largely undisturbed fauna. This patchiness may have been in part a response to a heterogeneous substrate. *Aulocystis* corallites show both reptant and erect branching morphologies, the more reptant corallites probably representing comparatively firmer substrates. This is supported by the relatively nonalate *Mucrospirifer* valves associated with the reptant clusters.

The increased faunal density within the fossil-rich layers indicates a greater oxygenation of the bottom waters than existed for the assemblages previously discussed. Although clustering indicates that transportation was not significant, there is evidence that current activity was greater. While storm winnowed shell hash layers were responsible for creating conditions favorable for colonization of the substrate, the convex up orientation of most brachiopod valves indicates continued current activity. The apparent resistance of *Athyris* to disarticulation has been reported by Bray (1971) who attributed it to a semi-infaunal life position. The inversion of *Pleurodictyum* coralla suggests that current disturbance may have been a major cause of mortality for this coral. Constriction and rejuvenesence of the coralla could also have resulted from the influx, or resuspension, of sediment associated with these currents. Episodic, low velocity currents seem to be indicated which were capable of flipping shells but unable to transport them. The *Aulocystis-Pleurodictyum* Assemblage, therefore, represents a moderately well oxygenated environment subject to relatively weak storm generated currents.

Crinoid-Bryozoan-*Athyris* Assemblage

Lithology and fossil content—A pseudoconcretionary carbonate bed, which may correlate with Grabau's (1899b) "*Nautilus* Bed" at the Lake Erie shore, (Kloc, 1983) overlies the *Aulocystis-Pleurodictyum* Assemblage. Unlike the ovoid, round bottomed concretions common in the Hamilton (Jordan, 1968), the limestone bodies of this bed are typically lenticular with a relatively uniform thickness of about 10 cm. The larger lenses assume the appearance of discontinuous limestone beds. Where they are closely spaced, the bioclastic wackestone to packstone lenses frequently grade laterally into a thin fossiliferous nodular layer.

As well as including the highest occurrences of *Pleurodictyum*, the pseudoconcretionary bed marks the beginning of a rich and diverse fauna that continues up to the "Trilobite Beds" (Grabau, 1899b). This bed is distinctive in the occurrence of large specimens of the bivalves *Modiomorpha*, *Ptychopteria*, and *Pseudoviculopecten* at the base of the lenses. Also, reaching its greatest abundance in this bed is the large spiriferid brachiopod *Spinocyrtia granulosa*. Epizoans are abundant, and *Spinocyrtia* is typically encrusted by *Aulocystis commensalis* and the trepostome bryozoan *Leptotrypella*. In addition, hederelloid bryozoans, spirorbid worm tubes, *Cornulites*, and crinoid holdfasts occur here and throughout the interval.

The 10 cm of shale overlying the pseudoconcretionary bed at Buffalo Creek possesses a unique assemblage of corals dominated by the large solitary rugosan *Heliophyllum halli*. Also present in this coral bed are large (about 15 cm diameter) fistuliporoid bryozoan mounds apparently associated with crinoid and bryozoan-rich layers (see Fig.6).

The shale between the coral bed and the "Trilobite Beds" contains abundant crinoids and bryozoans, and large brachiopods including *Athyris*, *Mediospirifer*, *Rhipidomella*, and *Megastrophia*. Bryozoans reach their greatest abundance at about the middle of this interval. Straight to gently sinuous pyritized burrows about 2 – 3 mm in diameter are common. This shale interval is characterized by thin, 1 – 4 cm thick fossiliferous layers in which crinoids, bryozoans, large brachiopods, and stereolasmid and aulocystid corals are concentrated. These alternate with much less fossiliferous layers often containing *Mucrospirifer* and *Zoophycos* spreiten.

Taphonomy — Regular alternations of fossil-rich and fossil-poor layers characterize the entire interval comprising the Crinoid-Bryozoan-*Athyris* Assemblage. Individual fossil-rich layers may be traceable over tens of meters, or may be very discontinuous and lenticular. The pseudoconcretionary bed at the base of this interval provides a detailed record of these layers as compaction is minimal and the preservation of skeletal material and traces is excellent. An acetate peel of this bed from Buffalo Creek revealed four layers which could properly be called sedimentation units and which contained distinct taphocoenoses (see Fig.7). A 2 cm thick crinoid packstone layer which overlies *Zoophycos*-burrowed micrite is divisible into two units. Its basal portion is reverse graded with fine crinoid and bryozoan material being succeeded by larger crinoid ossicles. In the upper portion the number of trilobites, small brachiopods, gastropods, and bryozoans increases dramatically. Many small brachiopods are articulated and geopetally filled. Above the packstone layer is an extensively burrowed unit from which *Zoophycos* is conspicuously absent. Large, frequently articulated brachiopods and unfragmented fenestrate and ramose bryozoans are common, and the micritic matrix contains very fine crinoid and bryozoan material. The sedimentary/faunal cycle then culminates in another *Zoophycos* layer.

At Cazenovia Creek this same bed similarly reveals alternating crinoid-bryozoan packstone and burrowed bryozoan wackestone units. In outcrop, stereolasmid corals may be found in life position projecting into the overlying shale with their apices in the fossil-rich layers. Similarly, Boardman (1960) found that most trepostome bryozoan zoaria originated in the thin fossiliferous beds.

As already suggested above, the taphonomy of this interval is quite complex. While some crinoid ossicles within the packstone layers are broken and abraded, many preserve details of their ornamentation, and stem segments several centimeters long are present. Similarly, bryozoans range from abraded fragments to essentially intact fenestrate colonies. Large brachiopods are typically disarticulated and convex up but also are found articulated and in life position. In addition, *Athyris*, *Rhipidomella*, *Mediospirifer*, and *Devonochonetes*, as well as the small solitary rugose coral *Stereolasma* frequently form loosely organized clusters ranging from about 10 to 40 cm in diameter.

Within the "coral bed" a twice geniculated corallum of a large *Heliophyllum halli* FORMA *irregulare* (Wells, 1937) was found (Fig.8). The morphology of this and similar, though less dramatic, specimens records the toppling and subsequent upward growth of the coral. Interestingly, this corallum was found with the calice inverted into the mud and a crinoid-bryozoan-rich layer deposited around the coral. Encrusting auloporid corals and crinoid holdfasts cover its upper surface. A broadly conical variant form of *H.halli*, FORMA *arachne* of Wells (1937), was frequently found inverted, reminiscent of *Pleurodictyum* in the assemblage below.

Environmental interpretation — The crinoid-bryozoan packstone layers described above are superficially similar to the "coquinites" described by Bowen and others (1974) in the Upper Devonian Sonyea Group of New York. Reverse grading is interpreted by them to be the result of the settling out of locally derived sediment resuspended by wave and bottom currents during severe storm activity, followed by the introduction of coarser material by "receding storm surges." Although this reconstruction is believed to be generally applicable to the packstone layers there are important differences in detail. One of these is the striking textural and faunal difference between the sediment above and below the packstone layer. Secondly, the unsorted and unfragmented fossils of the upper portion of the layer are clearly neither the products of transportation nor winnowing. This is supported by a study of disaggregated bedding plane samples from this layer which revealed an abundance of juvenile brachiopods indicating no loss due to winnowing or transportation (Miller, 1982).

Based upon this taphonomic information a probable sequence of events is outlined below. 1) Storm generated currents resuspend and winnow some fluid sediment, and fine skeletal material is redeposited forming a thin layer. 2) Storm surge currents deposit a layer composed dominantly, or solely, of transported crinoid debris. 3) A firmer substrate and abundant attachment sites permit the rapid colonization of the substrate by epifaunal suspension feeders including crinoids, bryozoans, and brachiopods. 4) With the gradual accumulation of mud and loss of attachment sites under relatively quiet water conditions, crinoids largely disappear and large brachiopods become established. 5) Continued mud deposition and bioturbation results in a fluid substrate gradually excluding the epifauna. This same basic process, with minor variations, is believed to account for all of the fossil-rich layers in the Crinoid-Bryozoan-*Athyris* Assemblage. In all cases it appears that the storm generated layers were critical to the development of the rich suspension feeding epifauna characteristic of this interval.

The storm current origin of the crinoid-bryozoan packstone layers is corroborated by the taphonomy of the *Heliophyllum* coralla within the coral bed. While the toppling and renewed

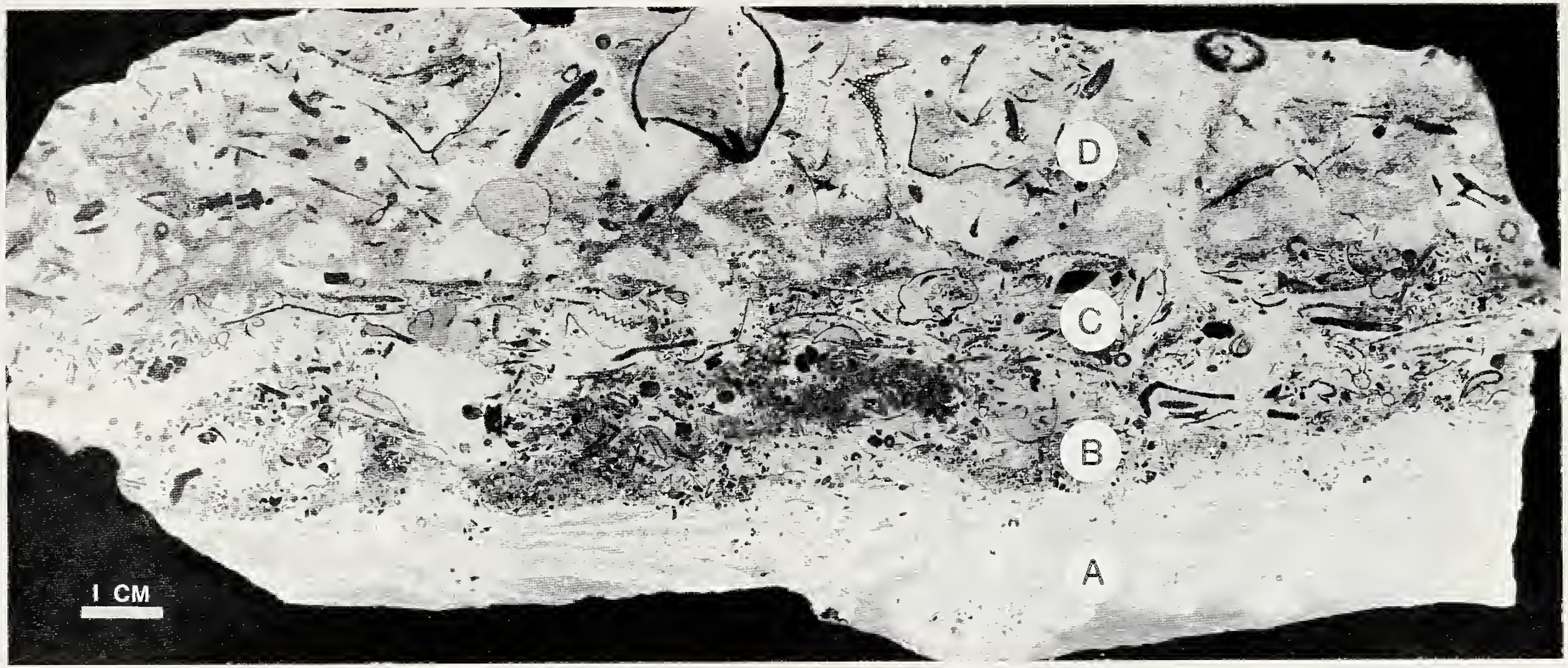


Figure 7 Photograph of portion of a limestone lense from Buffalo Creek made from an acetate peel. Four sedimentation units are distinguishable: A) *Zoophycos* layer, B) reverse graded crinoid layer, C) layer containing small brachiopods, gastropods, trilobites, and ramose bryozoans, D) extensively burrowed layer with large brachiopods and ramose and fenestrate bryozoans.

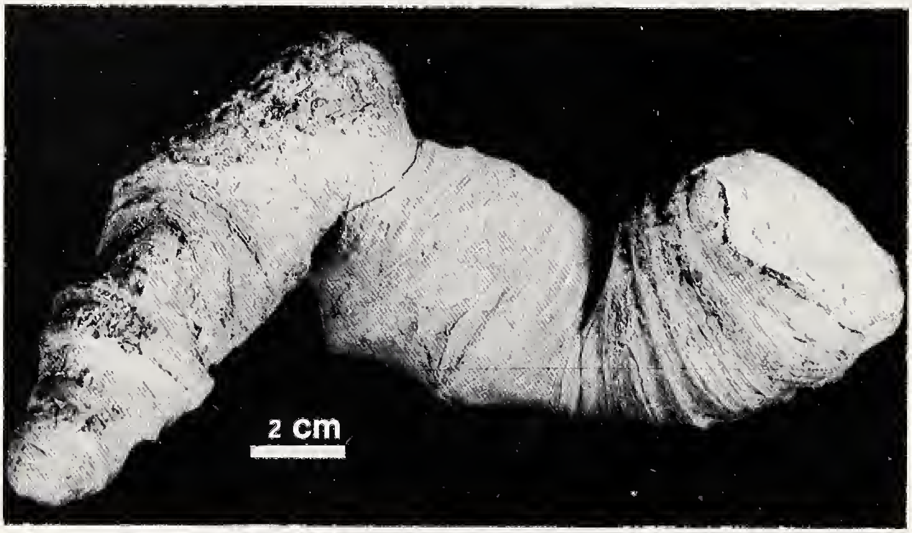


Figure 8 *Heliophyllum halli* **FORMA** *irregulare* **specimen showing final stable growth position resulting from three major episodes of toppling and renewed growth.**

growth of the large geniculated corallum shown in Figure 8 likely resulted from growth in a soft substrate, its last position was a very stable one which could have been disturbed only by a fairly strong turbulent current. The current which flipped this coral was probably also responsible for depositing the crinoid-bryozoan layer which surrounded it. The flipping of many smaller broadly conical coralla similarly reflects fairly energetic currents.

It is difficult to determine unambiguously the exact timing of burrowing activity within the proposed cycle of events. As mentioned by Fürsich (1978), *in situ* infaunal and epifaunal elements may occur in the same unit although they represent two distinct and temporally separated biocoenoses. The burrows which are preserved, therefore, may have largely post-dated colonization by the brachiopods and bryozoans occurring in the same sedimentation unit. Placing the activity of the *Zoophycos*-producing animal within this temporal sequence is especially difficult. As a systematic burrowing deposit feeder, it probably selectively exploited organic-rich layers (Simpson, 1970). Its apparent amensal relationship with epifauna may, therefore, be a consequence of concentrated feeding activity in organic-rich layers left uncolonized due to unfavorable substrate conditions. As described by Wetzel and Werner (1981), *Zoophycos* can be divided into vertical shaft spreiten, which are frequently disturbed by bioturbation, and horizontal feeding spreiten. The *Zoophycos* specimens are small, with feeding spreiten about 3 mm wide, at the lower size limit of those recorded by Wetzel and Werner. Accordingly, their unpreserved shaft spreiten were probably not long. Feeding spreiten occasionally disturb the crinoidal storm layers, but are absent from the bioturbated unit above the packstone. These observations suggest that the *Zoophycos* burrows were relatively shallow (10 cm or less) and can be assigned to later stages of the successional cycle.

The observed amensal relationships probably involve a complex set of sedimentological and biological factors. An adult filter feeding population may ingest spat, preventing their settlement. But as discussed by Eagle (1975), a suspension feeding fauna, such as initially colonized the storm deposit, cannot prevent the successful settlement of post-metamorphism juveniles. Passive secondary settlement by modern bivalves has been observed, and Eagle states that such settlement is "limited to hydrodynamically depositional environments where mud accumulates on the sea-bed." With the accumulation of mud, therefore, the fauna likely diversified and deposit feeders became established. The bioturbation of the sediment caused by deposit feeding activity may then have begun excluding sessile suspension feeding organisms. Rhoads and Young (1970) describe three causes of this exclusion: 1) clogging of filtering structures, 2) burying of newly settled larvae or discouraging their settlement, and 3) preventing attachment to an unstable substratum.

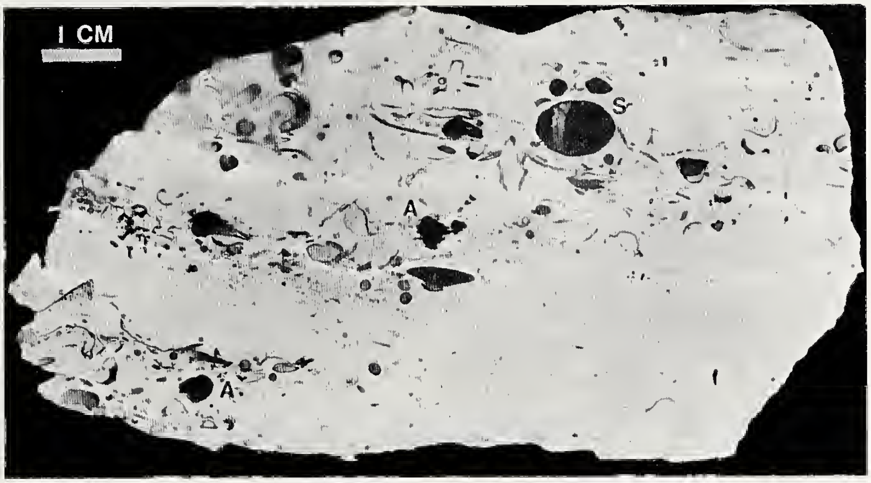


Figure 9 Photograph made from an acetate peel of the lowest Trilobite Bed at Cazenovia Creek showing thin regularly spaced fossil-rich layers containing Aulocystis (a) and Stereolasma (s).

The benthic environment in which the sediments of this assemblage were deposited was probably subject to major disturbance only during the most severe storms. At other times, the sea bottom, which was near wave base, experienced only gentle current action. The nature of the hydrologic regime helps to unravel the complex and seemingly conflicting taphonomy of this stratigraphic interval. Abraded crinoid ossicles and bryozoan fragments are the transported elements of the storm layers. The *in situ* fossil material includes flipped brachiopod valves and coral skeletons which record episodic low energy currents; and bryozoans, brachiopods, and corals preserved in life position by rapid burial during times of mud influx under very low energy conditions.

There is some evidence of depositional episodicity during the inter-storm periods. Abundant epizoans, lamellate fistuliporoid bryozoans, and occasional bored shells are suggestive of at least brief episodes of non-deposition. Evidence of episodic sediment influxes, on the other hand, is provided by the internal structure of large fistuliporoid bryozoan colonies. Figure 6 pictures an acetate peel of one hemispherical colony which shows regular layers of sediment overgrown by the bryozoan lamellae. It is only within these colonies that these, apparently frequent, sediment pulses were preserved from destruction by bioturbation.

During inter-storm periods the substrate was normally quite fluid. Most brachiopods have morphologies adapted to soft substrata such as the wide interareas in *Mediospirifer*, alate wings in *Mucrospirifer*, and the strongly concavo-convex shell forms in the strophomenids. A few spiriferid brachiopods were found with their "wings" pointed down into the sediment suggesting foundering into a soft unstable substrate. Diffuse-bounded burrows in the bioturbated unit of the limestone lenses also indicate high fluidity. Intense burrowing probably functioned to oxidize the sediment within a few centimeters of the sediment-water interface. Oxidation of organics within highly burrowed layers is supported by the petrography of the limestone lenses where pyrite is associated, not with the burrowed layer, but with the essentially unburrowed crinoid packstone layer. The decomposition of organics trapped in the storm layers may have been a major causal factor in the genesis of the carbonate lenses in a manner similar to that discussed above for the *Ambocoelia* concretions.

Aulocystis-Stereolasma Assemblage

Lithology and fossil content — The interval characterized by this assemblage includes the lowest Trilobite Bed of Grabau (1899b) and the shales immediately above and below. This bed varies from a bioclastic wackestone at Cazenovia Creek to a moderately fissile calcareous shale at Buffalo Creek. In addition to the trilobites *Phacops* and *Greenops*, this bed contains abundant *Aulocystis* and *Stereolasma*. *Ambocoelia* is common as is the gastropod *Naticonema*. This bed is also distinctive in the occurrence of the small strophomenid *Pholidostrophia*, the gastropod *Glyptotomaria*, and the crinoid *Ancyrocrinus* with its anchor-like holdfast. Impressions of the infaunal suspension feeder *Cypricardella* are common at Cazenovia Creek.

Taphonomy — Fossil-rich bands about 1 cm thick persist into the calcareous interval of the *Aulocystis-Stereolasma* Assemblage. However, crinoid layers are absent, and there is no evidence of transported skeletal material. Large brachiopods are also absent, except for *Rhipidomella*, and bryozoans are only a minor component of the fauna. *Aulocystis, Stereolasma, Ambocoelia*, and trilobites are the dominant constituents of the fossil-rich layers. These layers can be seen clearly in Figure 9 which shows an acetate peel of a Cazenovia Creek sample. Fossiliferous horizons are less clearly defined at Buffalo Creek where *Zoophycos* spreiten are somewhat more abundant. Vertical and horizontal burrows about 2 – 3 mm in diameter are present at both sites.

Preservation of skeletal material is excellent in this interval. At Cazenovia Creek articulated *Ambocoelia umbonata* shells occur in clusters about 10 cm in diameter, and are typically in a pedicle valve up position. Similarly, *Aulocystis* shows little or no fragmentation and occurs in clusters with *Stereolasma*, which occasionally is found in life position. Disarticulated trilobite remains are concentrated in these coral clusters; complete trilobites, often enrolled, also occur.

Environmental interpretation — The absence of large brachiopods and abundant bryozoans and crinoids is consistent with the conclusion made above that the crinoid "storm surge" deposits were critical to the development of a rich sessile epifauna. With the absence of many of the suspension feeders, an abundant supply of organic nutrients would have been made available to other organisms. This resource was utilized by a group of epifaunal deposit feeders including trilobites and the gastropods *Naticonema* and *Glyptotomaria* (Peel, 1978).

The excellent preservation of skeletal material shows that the fossil layers were not products of transportation, but surfaces of preferential colonization. Initial colonization of these surfaces was likely triggered by winnowing of sediment by relatively weak currents leaving a thin lag of fine skeletal material. The occurrence of brachiopods and corals in life position, and particularly the preservation of complete and enrolled trilobites, is strong evidence for rapid burial of the colonized substrates by mud under low energy conditions. These burial horizons are also believed to be genetically related to the early carbonate precipitation within this interval.

Generally low sediment influx, however, is suggested by the autecology of the taxa. According to Peel (1978), trochiform pleurotomariacean and platyceratid gastropods required a relatively firm bottom and clear non-turbid water. *Aulocystis* probably required similar conditions. Brett and others (1983) have also suggested that *Stereolasma* may have been more sensitive to sediment influx than its apparent environmental equivalent *Pleurodictyum*. A lower sedimentation rate also is consistent with the thinness of the interval and the close spacing of the fossiliferous horizons. The substrate was still fairly soft, however. Most stereolasmid corals have pointed apices indicating support by the sediment in an "iceberg" fashion, and some of these are twisted as a result of toppling and renewed growth. The peculiar anchor-like holdfasts of the crinoid *Ancyrocrinus bulbosus* were an adaptation for firm anchorage in a soft substrate (McIntosh and Schreiber, 1971). A generally quiet water, muddy but non-turbid environment with slow pulsational sedimentation and rare winnowing by weak currents is inferred.

Upper *Mucrospirifer-Devonochonetes* Assemblage

This assemblage is represented only by ten centimeters of shale at Cazenovia Creek and is absent from Buffalo Creek except for a single *Athyris*-rich horizon. The validity of this assemblage has been verified by recent detailed stratigraphic work by the author to the west where the section thickens. In addition to *Mucrospirifer* and *Devonochonetes*, this interval contains common *Athyris* and large *Ambocoelia umbonata* as well as *Mediospirifer*. *Athyris* is typically articulated and occasionally occurs in life position. The common presence of these brachiopods may reflect lower turbidity or a somewhat more oxygenated environment than normally existed in the Lower *Mucrospirifer-Devonochonetes* Assemblage.

Styliolina-Ostracod Assemblage

Lithology and fossil content — This uppermost interval includes two thin argillaceous limestone beds (i. e., the upper two "Trilobite Beds" of Grabau) and the associated shales. The assemblage is characterized by abundant ostracods and styliolinids, the ubiquitous brachiopods *Mucrospirifer* and *Devonochonetes*, and the trilobites *Phacops* and *Greenops*. The shale at the base of the interval contains several discontinuous bedding plane concentrations of ostracods, styliolinids, and brachiopods which were occasionally colonized by isolated aulocystid corals. The lower calcareous bed is extremely rich in styliolinids and approaches a grain supported texture. Burrowing is extensive with at least three distinct burrow types present: *Zoophycos* spreiten, large diffuse burrows about 4 – 10 mm in diameter, and sharply defined vertical burrows 2 – 3 mm in diameter. The upper bed is an argillaceous lime mudstone rich in *Styliolina* and very similar in texture to the "*Strophalosia* Bed." It is fairly homogeneous showing faint 1 1/2 – 2mm burrows and thin poorly defined concentrations of styliolinids.

Taphonomy — Nearly all brachiopods within this interval are disarticulated. *Mucrospirifer* valves show no consistent pattern of orientation from bed to bed or section to section, ranging from 1/3 to 2/3 convex up. However, at the top of the lower calcareous bed, *Mucrospirifer* occurs in small clusters of about six individuals, some of which are articulated and in life positions.

Trilobite material is typically highly disarticulated and patchily distributed. Complete trilobites do occur occasionally, particularly within the shales at the base of this interval.

Environmental interpretation – Like the *Styliolina-Ambocoelia* Assemblage, this interval is characterized by abundant styliolinids and small brachiopods. The concentration of styliolinids, especially in the lower calcareous bed, points to a very slow depositional rate with accumulation of the shells of these pelagic organisms. Disarticulation of the brachiopods is consistent with this interpretation. Long exposure of dead brachiopods at the sediment surface, where they would be subject to decomposition and reworking by scavengers, would result in the disarticulation and scattering of the valves.

Some pulsational deposition is still indicated. Remnants of thin styliolinid concentrations within the calcareous beds, and the more well-developed ostracod horizons in the lower shales, indicate significant periods of nondeposition. Preservation of complete trilobites, on the other hand, is clear evidence of rapid burial. The top of the lower calcareous bed with its clustered and articulated *Mucrospirifer* shells also appears to be a burial horizon. This particular horizon has an especially high concentration of styliolinids and is cut by sharp walled vertical burrows from above. The top of this bed is, therefore, interpreted as a buried firmground surface, which is consistent with the relatively non-alate morphology of the brachiopods.

The lower calcareous bed is extensively burrowed. The traces appear to represent two separate episodes of burrowing by two distinct groups of infauna superimposed on the same sediment. The first episode is represented by large diffuse-bounded burrows ranging from 4 to 10 mm in diameter in which fine skeletal material is concentrated. During this period the sediment was rather fluid. Darker vertical burrows which are mud filled and have sharp contacts characterize the second episode. At this time a relatively firm and partially dewatered substrate apparently existed. The absence of pyrite may be explained by the degree of bioturbation. Though the oxygen content of the water column was probably low and variable, constant mixing and irrigation of the sediment was sufficient to oxidize the organic material.

ENVIRONMENTAL SYNTHESIS

The basal three meters of the Wanakah Shale Member represent a regressional-transgressional event. The deepest water conditions were still within the photic zone but below storm wave base. The lower limit of the photic zone depends on a variety of factors including atmospheric conditions, latitude, and water clarity. Of interest here is the depth at which net primary production can occur. In clear tropical waters this zone extends to about 100 meters (Sumich, 1980) and, therefore, provides an estimate of the maximum depth. An estimate of the depth of storm wave base in the Hamilton Sea can be made assuming that modern low latitude shelf areas are good models. If so, a depth of about 40 meters would have been likely (Reineck and Singh, 1980). Below the influence of storms, the oxygen concentration and nutrient supply of the bottom water probably were both low and variable.

Shallowing of the sea brought the bottom through storm wave base to near normal wave base. Above wave base the water was well oxygenated and planktonic and organic food supplies were abundant. The environment within the sediment remained generally reducing, however, except where well bioturbated. The difference in depth required for these changes may have been relatively small. Depending on the frequency, duration, and intensity of storms and the geometry of the basin, the distance between normal wave base and storm wave base may have been small.

The fact that these various zones (i.e. below, near, and above wave base) are distinguishable in the fossil record indicates low depositional slopes. As pointed out by Anderson (1971), boundaries between zones are gradational and biodepositional zones must, therefore, be wide to be preserved as distinct entities. For depth related zones, such as the assemblages of this study, this implies a low slope. This is consistent with the existence of a broad carbonate shelf to the north.

SCALES OF STRATIGRAPHIC RESOLUTION

Most paleoecologic, non-reefal studies to date have been devoted to defining and interpreting large scale, biofacies patterns. A number of such studies have been done in the Devonian of the Appalachian basin in New York and elsewhere (Sutton, et al., 1970; Thayer, 1974; Bowen, et al., 1974; McGhee, 1976; Makurath, 1977; Feldman, 1980). As a consequence of the coarse scale of these investigations, much of the temporal and spatial resolution which is available from the fossil record is lost. Although general gradients and gross environmental changes are recognized, much valuable information about the depositional environment is bypassed. In an effort to retrieve this information, this study restricted its attention to a thin stratigraphic interval whose unique and variable fossil content permitted high stratigraphic resolution. Four scales of ecological/environmental processes, characterized by different durations, are evident within the three meters of shale and thin limestones investigated in this study. 1) The cumulative effects of very short-term ecological changes and depositional events, like those observed in modern marine ecosystems, are condensed within individual sedimentation units. 2) Sequences of sedimentation units define successional cycles initiated by storm events. 3) Assemblages, composed of contiguous beds with similar taphocoenoses, are arranged in a vertically symmetrical sequence recording a regressional-transgressional event through their taphonomic and sedimentological characteristics. 4) Changes in some faunal elements of the assemblages between the regressional and transgressional phases comprise the largest scale process.

The lower limit of resolution in this series is comparable to the upper limit of modern ecological studies. Sedimentation units are paleontologic time exposures (per: Schindel, 1980) and both spatial and temporal variation caused by short-term ecological processes are averaged out within them. Nonetheless, taxonomic abundance and size frequency distribution data (see Miller, 1982) can provide clues in the identification of the physical and biological factors which played important roles in determining short-term ecological changes. These include substrate stability and fluidity, availability of attachment sites, recruitment patterns, faunal ammensalism, and physical disturbance (Eagle, 1975; Peterson, 1977).

The repeated successional sequences characteristic of much of the central part of the stratigraphic interval would be "short-term successions" according to the terminology of Walker and Alberstadt (1975). They state that such successions usually occur where periodic physical disturbances destroy the existing fauna and a "rapid, biologically induced, successional reestablishment of original conditions" follows. In the successions discussed by Walker and Alberstadt, physical disturbance resulted in an unstable soft substrate which was stabilized by subsequent biologically controlled processes. The opposite trend, however, is exhibited in those described herein. Crinoidal storm deposits provided a firm substrate, rapidly colonized by epifauna, which was subsequently destabilized under normal depositional conditions by bioturbation. Related to the destabilization of the substrate was a progression from characteristic species to ubiquitous species and a concomitant decrease in species diversity. This finding is contrary to the classic understanding of successions on mud bottoms as put forward by R. G. Johnson (1972) and described in the fossil record by M. E. Johnson (1977) and others (see Walker and Alberstadt, 1975), in which pioneering opportunists prepare the substrate for other epifauna. The sudden appearance of a diverse group of suspension-feeding epifauna, however, does not exclude the likely possibility of an unresolved rapid faunal succession of the classic type during colonization of the storm deposit. After initial colonization, species diversity may have remained rather constant, or even increased, during the introduction of deposit feeding infauna. Continued slow sediment deposition and bioturbation gradually degraded the biocoenoses lowering diversity and restoring ubiquitous species to dominance.

In addition to the repeated short-term successions occurring within assemblages, the assemblages themselves occur in a regular cyclic sequence. This sequence is similar to Walker and Alberstadt's allogenic (i.e. physically induced) "long-term successions." Ferguson (1962), Hickey and Younker (1981), and Rollins and others (1979) have described sequences of paleoecologic units within stratigraphic intervals comparable in size to that of this study. All of these workers interpret the faunal changes in terms of transgressive and/or regressive events. In such cases, including the present study, sequences of assemblages are probably best discussed as facies changes, rather than ecological successions. This approach emphasizes their environmental significance and utility in defining depositional environments.

The assemblages are believed to reflect faunal belts which paralleled the carbonate shelf presumed to have bounded the basin to the north. Proceeding from the basin toward the shelf five belts are recognized: 1) a styliolinid and nautiloid belt with scattered concentrations of tiny ambocoelid brachiopods, 2) a *Mucrospirifer* and chonetid belt, 3) an auloporid coral belt with *Stereolasma* or *Pleurodictyum* and abundant trilobites, 4) a crinoid and bryozoan belt with a diverse group of large brachiopods, and 5) a coral belt characterized by large rugosans. These faunal belts are reflected very closely in the calcareous shales and thin limestones bounding the core Centerfield Limestone (Brett et al., 1983), and somewhat less closely in the shale above and below the Tichenor Limestone (Baird and Brett, 1981b). Both of these stratigraphic intervals of the New York Middle Devonian have been interpreted by Baird and Brett as regressive-transgressive events.

A regressive model for the highly fossiliferous calcareous horizons within the Hamilton is supported by this study. This model is at odds with the earlier transgressive model of McCave (1973) and the collapse of density stratification proposed by McCollum (1981) for the "*Pleurodictyum* zone." The primary argument for regression is the persuasive taphonomic and sedimentologic evidence for the increasing influence of storm activity toward the center of the roughly symmetrical section discussed in this paper. The regression model differs from the others in its assumption that the western part of the New York Devonian sea was influenced by a northern carbonate shelf rather than the prograding clastic shoreline to the east, or a shallowing of the basin to the west. McCollum reconstructs the lateral distribution of her faunal assemblage zones solely on the basis of an east-west, on-shore off-shore gradient. Her subsequent argument against a eustatic sea level drop is dependent on this reconstruction. McCave's model attempts to correlate calcareous horizons within the Hamilton to transgressive pulses onto the generally prograding Catskill clastic shoreline. However, this source of clastic input is too far distant to have had a controlling influence, especially considering the likely existence of a depositional trough in the Finger Lakes area.

The longest recorded ecologic process is the apparent replacement of some taxa between the regressive and transgressive episodes. *Pleurodictyum* is replaced by *Stereolasma* and *Spinocyrtia* is replaced by the other large brachiopods *Athyris*, *Mediospirifer*, and *Rhipidomella*. In these examples no new taxa are introduced, but existing taxa increase in abundance, perhaps to fill vacated niche space. Several taxa appear for the first time in the trangressive *Aulocystis-Stereolasma* biosome, including *Ancyrocrinus*, *Pleurotomaria*, and *Pholidostrophia*. Two possible explanations for these observations are put forward here. The transgressive portion of the section is significantly thinner than the regressive portion, and this asymmetry may indicate a considerable difference in the length of time over which these events occurred. Under these circumstances, deeper-water faunal belts would have replaced shallower-water ones more rapidly than shallow water faunas would have invaded deep water zones. In addition, transgressive and regressive regimes would have placed the fauna under fundamentally different stresses as pointed out by Rollins and others (1979). During transgression organisms are under primarily allogenic stress as they colonize new substrates, but during regression they are under both allogenic and autogenic stress as established habitats are restricted and invaded. The taxonomic asymmetries may, therefore, be a result of these stress differences resulting from the rate and direction of environmental change. The other alternative involves changes in physical parameters of the environment. The assymetry of the section probably was at least partially the consequence of different depositional rates. Brett and others (1983) have proposed that the *Pleurodictyum* coral association may have been more tolerant of turbidity and higher sedimentation rates than the *Stereolasma* association which appears to have occupied an analogous bathymetric onshore-offshore position. Supporting this idea is the greater thickness of the regressive, *Pleurodictyum* dominated, half cycle. Also environmental factors such as source area, current patterns, and climate may have changed such that the transgressive and regressive assemblages were characterized by somewhat different conditions of water turbulence, oxygen content, temperature, or resource supply.

CONCLUSIONS

This paleoecological study has recovered valuable faunal, taphonomic, and sedimentological information which has enabled the depositional environment of the "*Pleurodictyum* zone" to be reconstructed in detail. As a result, conclusions can be drawn which have applications to the Hamilton shales in particular and to marine muddy bottom benthic biocoenoses in general. 1) A regressive model for the highly fossiliferous limestone units of the Hamilton Group is supported by taxonomic, taphonomic, and sedimentological evidence. 2) Assemblages, comprised of stratigraphically contiguous taphocoenoses, can be defined by presence-absence data and reflect faunal belts paralleling a northern carbonate shelf. 3) A hierarchy of ecological/environmental processes ranging from short-term successions to regressive-transgressive events, can be resolved from fine-grained rocks within a vertical interval of only a few meters. 4) A critical factor for sessile epifaunal benthic organisms on a muddy substrate is the availability of attachment surfaces. For this reason, storm deposits were the key to the development of the rich and diverse epifaunal suspension-feeding assemblages which characterize the "*Pleurodictyum* Zone."

ACKNOWLEDGMENTS

I express my sincere appreciation to Dr. James R. Beerbower, who suggested and supervised the research from which this article was drawn. The field work was supported by a research grant (No. 2694-80) from the Geological Society of America.

The critical reviews of G. C. McIntosh, D. L. Meyer, and J. C. Brower were of great help in improving my clarity of thought and presentation.

REFERENCES CITED

Anderson, E.J. 1971. Environmental models for Paleozoic communities. Lethaia 4:287-302.

Baird, G.C. 1979. Sedimentary relationships of Portland Point and associated Middle Devonian rocks in central and western New York. New York State Mus. Bull. 433, 24 p.

__________ 1981. Submarine erosion on a gentle paleoslope: a study of two discontinuities in the New York Devonian. Lethaia 14:105-122.

__________ and Brett, C.E. 1981a. Stop descriptions: Stops 1, 5, 8, 9, and 10. *In* Oliver, W.A., Jr. and Klapper, G.A. (eds.), Devonian Biostratigraphy of New York. Intl. Union Geol. Sci., Subcomm. on Devonian Strat., Washington, D.C., p. 3-37.

__________ and Brett, C.E. 1981b. Submarine discontinuities and sedimentary condensation in the upper Hamilton Group (Middle Devonian): examination of marine shelf and paleoslope deposits in the Cayuga Valley. New York State Geol. Assn. Guidebook, 53rd Ann. Mtg., SUNY at Binghamton, p. 115-145.

Berner, R.A. 1968. Calcium carbonate concretions formed by the decomposition of organic matter. Science 159:195-197.

__________ 1970. Sedimentary pyrite formation. Amer. J. Sci. 268:1-23.

Bray, R.G. 1971. Ecology and "life history" of Mid-Devonian brachiopod clusters, Erie County, New York. Doctoral dissertation, McMaster University, 162 p.

Brett, C.E., Baird, G.C., Gray, L.M., and Savarese, M.L. 1983. Middle Devonian (Givetian) coral associations of western and central New York State. *In* Silurian and Devonian Corals and Stromatoporoids of New York. 4th Intl. Symp. Fossil Cnidaria. p. 65-107.

Boardman, R.S. 1960. Trepostomatous bryozoa of the Hamilton Group of New York State. U.S. Geol. Survey Prof. Paper 340, 87 p.

Bowen, Z.P., Rhoads, D.C., and McAlester, A.L. 1974. Marine benthic communities in the Upper Devonian of New York. Lethaia 7:93-120.

Buehler, E.J. and Tesmer, I.H. 1963. Geology of Erie County, New York. Buffalo Soc. Nat. Sci. Bull., Vol. 21, No. 3, 118 p.

Cheetham, A.H. and Hazel, J.E. 1969. Binary (presence-absence) similarity coefficients. J. Paleont. 43:1130-1136.

Cleland, H.F. 1903. A study of the fauna of the Hamilton Formation of the Cayuga Lake Section in central New York. U.S. Geol. Survey Bull. 206, 111 p.

Cooper, G.A. 1930. Stratigraphy of the Hamilton Group of New York. Amer. J. Sci. 5(19):116-134, 214-236.

Davis, J.C. 1973. Statistics and Data Analysis in Geology. John Wiley & Sons, New York, 550 p.

Dick, V.B. 1982. Taphonomy and depositional environments of Middle Devonian (Hamilton Group) pyritic fossil beds. Master's thesis, University of Rochester, 118 p.

Eagle, R.A. 1975. Natural fluctuations in a soft bottom benthic community. J. Marine Biol. Assn. U.K. 55:865-878.

Feldman, H.R. 1980. Level-bottom brachiopod communities in the Middle Devonian of New York. Lethaia 13:27-46.

Ferguson, L. 1962. The paleoecology of a lower Carboniferous marine transgression. J. Paleont. 36:1090-1107.

Fürsich, F.T. 1978. The influence of faunal condensation and mixing on the preservation of fossil benthic communities. Lethaia 11:243-250.

Grabau, A. 1899a. The geology of Eighteen Mile Creek and the lake shore sections of Erie County, New York. Buffalo Soc. Nat. Sci. Bull. 6(1).

__________ 1899b. The palaeontology of Eighteen Mile Creek and the lake shore sections of Erie County, New York. Buffalo Soc. Nat. Sci. Bull. 6(2,3,4).

Hickey, D.R. and Younker, J.L. 1981. Structure and composition of a Pennsylvanian benthic community. J. Paleont. 55:1-12.

Johnson, M.E. 1977. Succession and replacement in the development of Silurian brachiopod populations. Lethaia 10:83-93.

Johnson, R.G. 1964. The community approach to paleoecology. *In* Imbrie, J. and Newell, N.D., eds., Approaches to Paleoecology. John Wiley & Sons, New York, 107-134.

__________ 1972. Conceptual models of benthic marine communities. *In* Schopf, T. J., ed., Models in Paleobiology. Freeman, Cooper, and Co., San Francisco, 148-159.

Jordan, F.W. 1968. Genesis of carbonate concretions in the upper Ludlowville, Middle Devonian of Erie County, New York. Master's thesis, McMaster University, 66 p.

Kaesler, R.L. 1970. The cophenetic correlation coefficient in paleoecology. Geol. Soc. Amer. Bull. 81:1261-1266.

Kauffman, E.G. 1873. Cretaceous assemblages, communities, and associations. Western Interior United States and Caribbean Islands. *In* Ziegler, A.M. and others, Principles of Benthic Community Analysis. Notes for a Short Course, Miami, University of Miami.

Kloc, G.J. 1983. Stratigraphic distribution of ammonoids from the Middle Devonian Ludlowville Formation in New York. Master's thesis, SUNY at Buffalo, 78 p.

Lawrence, D.R. 1968. Taphonomy and information losses in fossil communities. Geol. Soc. Amer. Bull. 79:1315-1330.

MacDonald, K.B. 1975. Quantitative community analysis: recurrent group and cluster techniques applied to the fauna of the Upper Devonian Sonyea Group, New York. J. Geol. 82:473499.

Makurath, J.H. 1977. Marine faunal assemblages in the Silurian-Devonian Keyser Limestone of the central Appalachians. Lethaia 10:235-256.

McCave, I.N. 1973. The sedimentology of a transgression: Portland Point and Cooksburg Members (Middle Devonian) New York State: J. Sediment. Petrol. 48:484-504.

McCollum, L.B. 1981. Distribution of marine faunal assemblages in a Middle Devonian stratified basin, lower Ludlowville Formation, New York. Doctoral dissertation, SUNY at Binghamton, 143 p.

McGhee, G.R. 1976. Late Devonian benthic marine communities of the central Appalachian Allegheny Front. Lethaia 9:111-136.

McIntosh, G. and Schreiber, R.L. 1971. Morphology and taxonomy of the Middle Devonian crinoid *Ancyrocrinus bulbous* Hall 1862. Univ. of Mich. Mus. Paleont. Contrib. 5(23):381-403.

Mello, J.F. and Buzas, M.A. 1968. An application of cluster analysis as a method of determining biofacies. J. Paleont. 42:747-758.

Millendorf, S.A., Srivastava, G.S., Dyman, T.A., and Brower, J.C. 1978. A Fortran program for calculating binary similarity coefficients. Computers and Geosciences 4:307-311.

Miller, K.B. 1982. Paleomicroecologic analysis of depositional environments and sequences, "*Pleurodictyum* Zone," Ludlowville Formation, Middle Devonian of western New York. Master's thesis, SUNY at Binghamton, 131 p.

Oliver, W.A., Jr. 1976. Biogeography of Devonian rugose corals. J. Paleont. 50:365-373.

Palmer, T.J. and Palmer, C.D. 1977. Faunal distribution and colonization strategy in a Middle Devonian hardground community. Lethaia 10:179-199.

Peel, J.S. 1978. Faunal succession and mode of life of Silurian gastropods in the Arisaig Group, Nova Scotia. Paleontology 21:285-306.

Peterson, C.H. 1977. The paleoecological significance of undetected short-term temporal variability. J. Paleont. 51:976-981.

Reineck, H.E. and Singh, I.B. 1980. Depositional sedimentary environments. Springer-Verlag, New York, 549 p.

Rhoads, D.C. 1970. Mass properties, stability, and ecology of marine muds related to burrowing activity. *In* Crimes, T.P. and Harper, J.C., eds., Trace Fossils. Geol. J. Spec. Issue 3:391-406.

__________ and Young, D.K. 1970. The influence of deposit-feeding organisms on sediment stability and community trophic structure. Marine Research 28:150-177.

Rollins, H.B., Carothers, M., and Donahue, J. 1979. Transgression, regression and fossil community succession. Lethaia 12:89-104.

Schindel, D.E. 1980. Microstratigraphic sampling and the limits of paleontologic resolution. Paleobiology 6:408-426.

Simpson, S. 1970. Notes on *Zoophycos* and *Spirophyton*. *In* Crimes, T.P. and Harper, J.C., eds., Trace Fossils. Geol. J. Spec. Issue 3:505-514.

Stanton, R.J., Jr. 1976. Relationship of fossil communities to original communities of living organisms. *In* Scott, R.W. and West, R.R., eds., Structure and Classifiction of Paleocommunities. Dowden, Hutchinson & Ross, Inc., Stroudsburg, PA, p. 107-142.

Sumich, J.L. 1980. An Introduction to the Biology of Marine Life, 2nd ed. Wm. C. Brown Co., Dubuque, Iowa, 359 p.

Sutton, R.G., Bowen, Z.P., and McAlester, A.L. 1970. Marine shelf environments of the Upper Devonian Sonyea Group of New York. Geol. Soc. Amer. Bull. 81:2975-2992.

Thayer, C.W. 1974. Marine paleoecology in the Upper Devonian of New York. Lethaia 7:121-155.

Walker, K.R. and Alberstadt, L.P. 1975. Ecological succession as an aspect of structure in fossil communities. Paleobiology 1:238-257.

Wells, J.W. 1937. Individual variation in the rugose coral species *Heliophyllum halli* E. & H. Palaeontographica Americana 2:6.

Wetzel, A. and Werner, F. 1980/81. Morphology and ecological significance of *Zoophycos* in deep-sea sediments off NW Africa. Palaeogeog. Palaeoclim. Palaeoecol. 32:185-212.

Ziegler, A.M. 1973. The community technique. *In* Ziegler, A.M. and others, Principles of Benthic Community Analysis: Notes for a Short Course, Miami, University of Miami.

Deposition and Early Diagenesis of a Middle Devonian Marine Shale: Ludlowville Formation, Western New York

Glenn T. Wygant[1]
Department of Geological and Environmental Sciences
State University of New York at Binghamton

ABSTRACT

Petrographic study and paleoecologic analysis of the Ledyard and Wanakah Shale Members (Ludlowville Formation, Middle Devonian) in western New York reveal three subdivisions, each with a variety of depositional and early-diagenetic features. The lower and middle Ledyard contain few shelly fossils and burrows, high concentrations of disseminated organic material, and a relatively large abundance of primary pore-filling pyrite and organic-walled microfossils. The upper Ledyard and the lower Wanakah contain many shelly fossils and burrows and relatively few disseminated organic particles and organic-walled microfossils. Lastly, the middle and upper Wanakah contain a moderate abundance of shelly fossils and burrows and a somewhat greater abundance of disseminated organic material and organic-walled microfossils than the second subdivision.

The Ledyard and Wanakah Shale Members had a complex depositional and early-diagenetic history. The lower subdivision represents deposition on a poorly-oxygenated (anoxic-?)substrate. Consequently, benthic animals could only rarely inhabit these substrates and bioturbation was minimal. This resulted in the preservation of more organic material that, in turn, resulted in more primary pore-filling pyrite. The upper two subdivisions represent deposition on a well-oxygenated substrate; the middle subdivision represents optimal conditions for benthic organisms. Most of the sediments in all three subdivisions are argillaceous micropellets that were formed by suspension-feeding animals: either pelagic or softbodied epibenthic forms.

INTRODUCTION

The Ledyard and Wanakah Shale Members of the Ludlowville Formation (Hamilton Group, Middle Devonian) consist of clayshales, concretionary beds, and thin limestone beds and include a variety of fossil assemblages. These members were studied at exposures in the bed and banks of Cazenovia Creek near Spring Brook, New York (Fig. 1).

The objective of this study was to interpret the depositional and early-diagenetic conditions involved in the genesis of the Ledyard and Wanakah Shale Members. These interpretations are based on paleontologic, sedimentologic, and early-diagenetic features observed at the outcrop and in thin section.

Previous studies of the Ledyard and Wanakah Shale Members in western New York have dealt with stratigraphy (Cooper, 1930; Rickard, 1975; Baird, 1979), systematic paleontology (Hall, 1857; Grabau, 1899; and numerous others), and paleoecology (most recently: McCollum, 1980; Miller, 1982). Petrographic and sedimentologic studies of the Ludlowville Formation have been restricted to the near-shore and nonmarine strata in central New York (see McCave, 1967 for a reference list). The two notable exceptions are studies of marine shale by Towe (1961) and Jordan (1968), but petrographic interpretations were not the main emphasis of these studies.

GENERAL STRATIGRAPHIC AND SEDIMENTOLOGIC FRAMEWORK

The Hamilton Group in New York is an eastward thickening clastic wedge that serves as the hypostratotype for Middle Devonian rocks in the United States. The group crops out along an east-west trending belt from Lake Erie to the Hudson River through southeastern New York and into eastern and central Pennsylvania. At Lake Erie, the Hamilton Group consists of 80 meters of marine shale and limestones. Eastwards, the group thickens to 920 meters of alluvial sandstones (near the Hudson River).

The clastic sediments within the Hamilton Group were deposited as a tectonic delta complex (Friedman and Johnson, 1966). The delta complex in the northern portion of the Appalachian Basin underwent minor transgressive and regressive sedimentation cycles, perhaps in response to changes in sea level (McCave, 1967). These cycles also are recognizable

---

[1] Present address: Edwards and Kelcey, Inc.

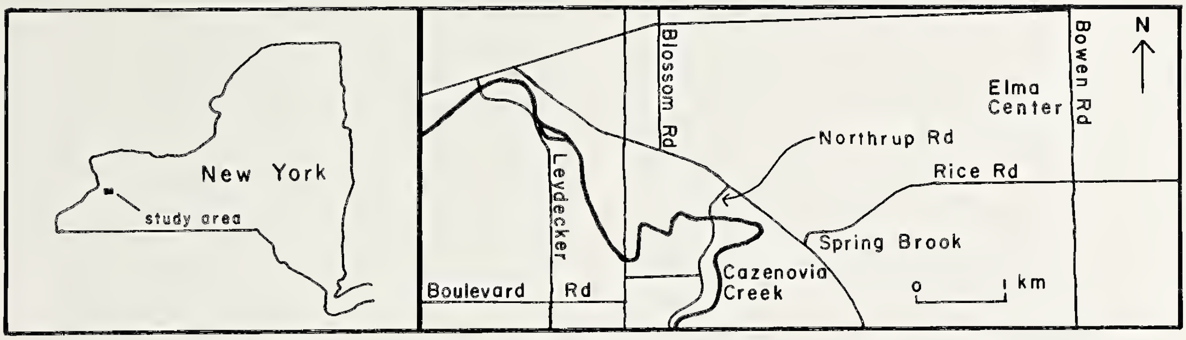


Figure 1. Location of study area on Cazenovia Creek, approximately 16 kilometers southeast of Buffalo, New York (modified from Buehler and Tesmer, 1963).

within the basinal deposits; black shales represent transgressive maxima and sandstones or carbonates cap regressive sequences (Brett and others, 1983; Baird and Brett, 1983). Alternatively, the black shales might have resulted from a restriction in the vertical mixing of seawater, possibly by the formation of a pycnocline (McCollum, 1980).

The Hamilton Group consists of four formations (Rickard, 1975): the Marcellus (lowest), the Skaneateles, the Ludlowville, and the Moscow (uppermost). In western New York the Ludlowville Formation consists of marine shales and limestones, and is subdivided into the Centerfield Limestone Member (basal), the Ledyard Shale Member, and the Wanakah Shale Member (uppermost).

The presently accepted stratigraphic correlations within the Ludlowville Formation are those suggested by Cooper (1930) as refined by Rickard (1975) and Baird (1979). Only recently have the shale facies relationships been studied in detail (Baird, 1979; 1981) and they have been found to be more complex than previously envisioned (Fig. 2). The Ledyard and Wanakah Members change gradually from gray shale in western New York to black shale in central New York (Baird, 1979; McCollum, 1980; Baird and Brett, 1981, 1983). This black shale is interpreted to have been deposited in a deep-water, less-oxygenated environment. Further east, the black shale passes into fossiliferous shales and siltstones of the King Ferry Member.

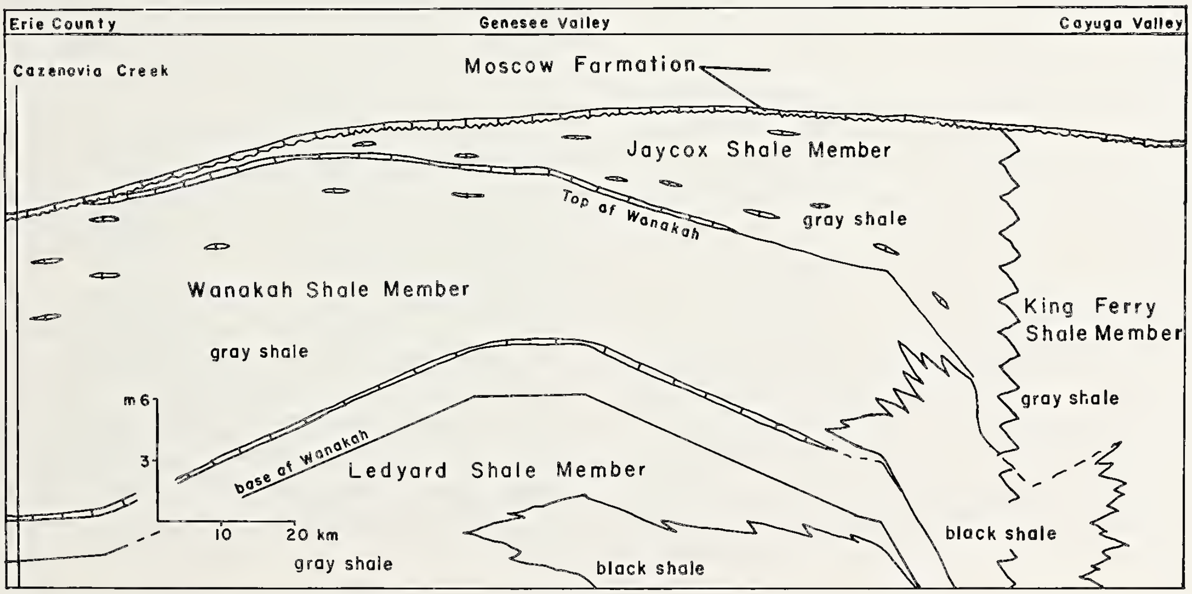


Figure 2. Presently accepted stratigraphic relationship of Ludlowville shales in western and central New York (modified from Baird, 1979).

Recent studies have interpreted the depositional settings of stratigraphic horizons within the Ludlowville Formation. The lower portion of the Wanakah Member (the *Pleurodictyum* Zone) apparently represents sedimentation in shallow water where occasional storms reworked the sediments (Miller, 1982, 1983). Dick and Brett (1982) have postulated that the pyrite beds (pyritized fossils and pyrite nodules) within the Hamilton Group are the result of episodic, burial events in anoxic, non-sulfidic environments. Studies of concretionary beds within the Ludlowville Formation by Jordan (1968), Beerbower (1981), and Speyer and Brett (1983) have concluded that such beds also were produced by rapid deposition.

The burial history of the Ludlowville Formation in western New York is poorly understood. However, the thermal maturation of the Hamilton Group in southwestern Ontario (approximately 20 km northwest of the region studied here) has been documented by Legall and others (1981). They have concluded, based on the color alterations of conodonts and acritarchs, that the formation reached a maximum temperature of only 60°C during burial.

A SUMMARY OF PROCESSES AND PRODUCTS IN GENESIS OF MARINE SHELF AND BASIN SHALES

The following general discussion concerns the deposition and diagenesis of fine-grained sediments and serves as the basis for interpretations presented in this study. A classification, or model, that can be used in interpreting the sedimentary environments of all fine-grained sediments has been proposed by Berner (1981). Berner's classification relates the thermodynamic stability of authigenic minerals to the life processes occurring within a sediment. The sedimentary environments are initially subdivided into oxic and anoxic. Oxic environments are characterized by authigenic iron oxides. Anoxic environments are further subdivided into sulfidic and nonsulfidic (post-sulfidic if sulfate was depleted), that are characterized respectively by iron sulfides and siderite. The concentration of dissolved oxygen and dissolved sulfate can become progressively depleted during burial (because they are utilized by organisms), so a layer of sediment can be subjected to all three environments.

Fine-grained particles suspended in seawater settle out of suspension as they are combined into larger and faster settling aggregates. One such aggregation process is flocculation — the electrostatic attraction of clay minerals. Flocculation is most common in river-fed estuaries because the electrostatic attraction of suspended clay minerals increases as the salinity of their fluid medium increases (Whitehouse and others, 1960). The force of the electrostatic attraction varies among the clay minerals, resulting in a depositional segregation of different clay minerals perpendicular to the mouths of rivers (Edzwarld and O'Melia, 1975; Gibbs, 1977). Another process which aggregates suspended sediment is the production of fecal and pseudofecal pellets by suspension-feeding animals. These benthic and pelagic animals are responsible for the deposition of fine-grained sediments within many modern sedimentary environments (Verwey, 1952; Schrader, 1971; Smayda, 1971; Pryor, 1975; Syvitski and Lewis, 1980). Pelletization of suspended sediment by benthic animals was common within ancient sedimentary environments (Wilson, 1975) but the importance of pelagic animals in depositing sediment within ancient environments is less clear.

Bioturbation greatly affects the physical and chemical characteristics of a substrate (Rhoads, 1974; Aller, 1977). This mechanical mixing of sediments destroys the depositional textures and moves seawater through the sediments much faster than by diffusion alone. Thus bioturbation increases the porosity of muddy substrates; intense bioturbation can result in porosities exceeding 60 percent (Rhoads and Young, 1970). Resuspension of sediment by only weak currents is then possible and, in turn, the turbidity so generated is detrimental to many benthic suspension feeders (Rhoads and Young, 1970). The bioturbation of sediments decreases and eventually ceases with depth (or burial) because oxygen becomes depleted from the pore water. Some large burrowers, however, live in oxygen-depleted sediments and obtain oxygenated water from above (Aller, 1977). As a result, the boundary between oxic and anoxic pore water is gradational.

The abundance of organic tissues and molecules within anoxic sediment is important because the anaerobic decomposition of this material has a profound influence on the chemistry of the pore water. Factors that control the amount of organic material reaching anoxic sediment are: biological productivity in the overlying seawater, the abundance of heterotrophic animals living on and within the sediment (while oxygenated), and the rate of sedimentation (Heath and others, 1977). The organic material within the anoxic sediment is decomposed by sulfate-reducing bacteria and then later by fermenting microorganisms when dissolved sulfate becomes depleted (Curtis, 1980).

Iron sulfides are a byproduct of the decomposition of organic material by sulfate-reducing bacteria and are quite common within fine-grained, marine sediments (Berner, 1970). The formation of iron sulfides is not only dependent upon sufficient amounts of dissolved sulfate and labile, organic material, but also upon the presence of mobile, ferrous iron. This ferrous iron can be supplied by clay minerals (Drever, 1971) and iron-oxide coatings on detrital grains.

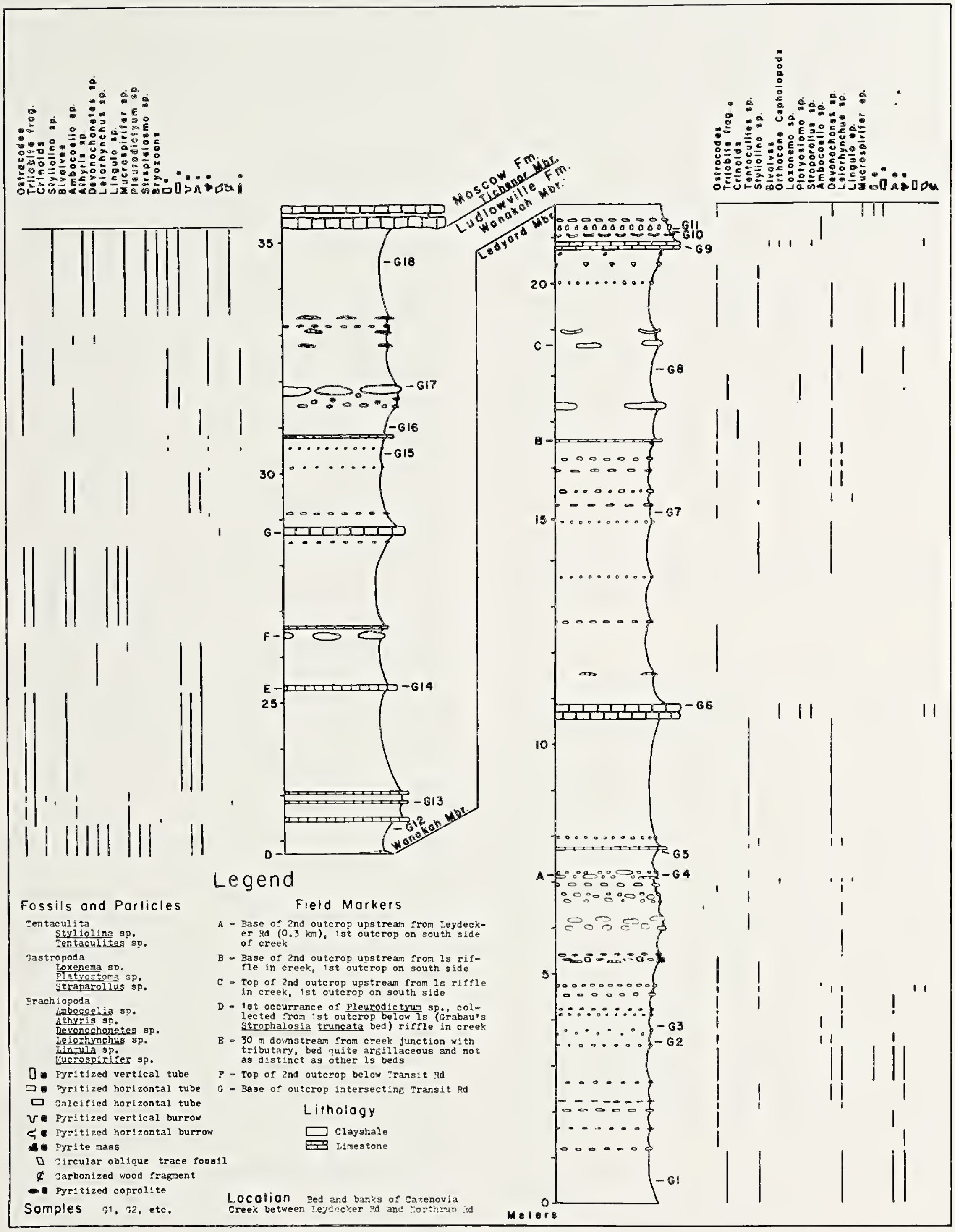


Figure 3. Composite measured section.

Generalizations concerning the pH of the pore water within the modern sulfidic environments are difficult because of the various chemical reactions that can occur. For instance, the pyrite reaction itself is acidic but some closely associated reactions are basic (e.g., ammonia is produced by the anaerobic decomposition of proteinaceous compounds). Petrographic and geochemical studies of the pyrite formed within ancient sulfidic microenvironments have revealed that the chemical conditions also were quite variable (Dick and Brett, 1982; Hudson, 1982; Raiswell, 1982).

Mineralogical and chemical changes occur during burial in response to increases in the ambient temperature. One such change involves mixed-layered illite/smectite, that progressively changes from less than 20 percent to approximately 80 percent illite layers in response to the temperature increase (Hower and others, 1976). The decomposition of organic material during burial in the post-sulfidic environment occurs first by a variety of temperature-independent processes and then later by a temperature-dependent process termed catagenesis (Hunt, 1979, p. 101). The thermal alterations of organic material and mixed-layered illite/smectite are irreversible; hence they are used to characterize the burial history of a sediment (Gutjahr, 1966; Hower and others, 1976).

LITHOLOGY AND PALEOBIOLOGIC SEQUENCE OF THE WANAKAH AND LEDYARD SHALES

Lithology

The section along Cazenovia Creek consists of shale and occasional beds of argillaceous limestone and calcium carbonate concretions (Fig. 3). All of the shales are clayshales; they are composed of more than 67 percent clay-sized particles (Lundegard and Samuels, 1980). In the field the unweathered clayshales are medium gray (2.5Y N5/0) in color, while the weathered clayshales are light gray (2.5Y N7/0). It probably is because of their light-gray color that the Ludlowville shales at Cazenovia Creek have been termed calcareous ever since the initial work of Grabau (1899). Most of the limestones are wackestones and packstones, with styliolinids being the most common fossil. The two thickest limestone beds display a sharp and irregular contact with the adjacent clayshale. Portions of these limestones protrude into the neighboring clayshale and resemble the tops and bottoms of concretions. All of the weathered limestones and clayshales possess imperfect fissility and this is perhaps suggestive of bioturbation (Beerbower and others, 1969; Byers, 1974).

Calcium-carbonate concretions within a bed share similar morphologies (i.e., shape and size), are consistent in their lateral spacing (i.e., distance between adjacent concretions), and have distinct fossil or lithological associations (Fig. 3). In many concretionary beds, molds of molluscan shells are present within the concretions and the laterally adjacent clayshale. Also associated with some concretionary beds are lenses of argillaceous limestone (clayshale with micrite cement). Adjacent to and even encased within the concretions in several beds are pyritized trail-like traces and pyrite nodules. These nodules are up to 8 cm in diameter and frequently contain inclusions of clayshale. A few concretionary beds are so closely spaced that it was found to be more appropriate to refer to these as concretionary zones (Fig. 3: 5.3 meters, 7.1 meters, and 31.7 meters above the lowest exposure of Ledyard shale).

Fossils

Many shelly fossils are contained within portions of these clayshales and argillaceous limestones. Most numerous are brachiopods, bryozoans, corals, echinoderms, molluscs, and trilobites. The stratigraphic distribution of these fossils varies considerably, as shown in Figure 3.

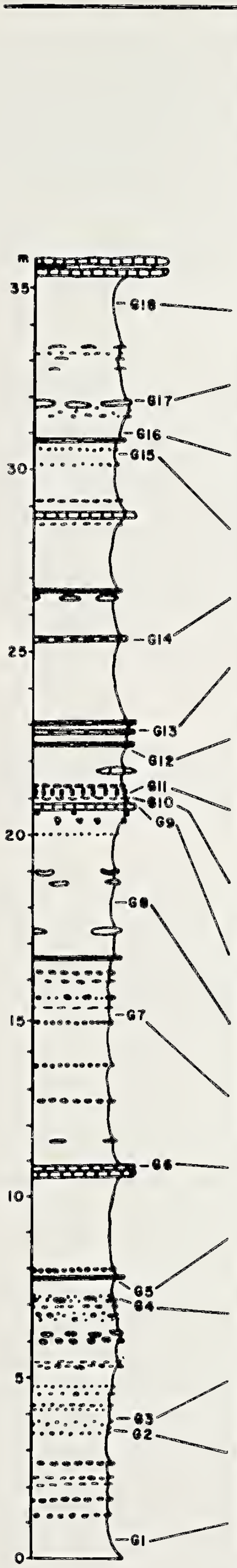
The preservation of these shelly fossils also varies within the stratigraphic section. Some beds contain brachiopods and corals that are intact and situated in life position. In other beds the skeletal material is fragmented (Beerbower and others, 1969; Miller, 1982). Mineralogically, most of the skeletal material is, and originally was, composed of low-magnesium calcite; these fossils are preserved in exquisite detail. The skeletal material of the nonstyliolinid molluscs originally consisted of aragonite and has been dissolved. These molluscs are preserved only as external molds and sediment casts, though on a few bedding surfaces the molluscan aragonite was replaced by pyrite.

Trace fossils are also present and occur in three varieties: pyritized trail-like traces, pyritized tubes, and obliquely inclined cylinders. The stratigraphic distribution of these traces is shown in Figure 3. Although none of these represent the more distinctive trace fossil taxa, their presence has paleoecological significance.

Pyritized trail-like traces are both horizontal and inclined but horizontal traces are more common. Individual traces vary from straight to meandering and can be followed for several centimeters. In cross section they are oval, having a maximum dimension of about 2 mm. The traces are interpreted as being locomotion traces (Simpson, 1975) made by worms burrowing beneath the sediment-water interface. The resulting voids, perhaps kept open with mucilage (Carney, 1982), were filled with iron sulfides before any significant compaction occurred.

Pyritized tubes are restricted in distribution to a few concretionary beds, occurring on the exterior of concretions and in the laterally adjacent clayshale. A typical tube is 1 cm wide and 3 cm long and has distinct, pyritized walls. Elongate crystals of pyrite are oriented towards the center of the tubes and completely fill the chambers. If any ornamentation was present on the exterior of the walls it was obliterated by pyritization. These tubes are interpreted as having been dwelling structures.

Obliquely inclined cylinders are found in the argillaceous limestones. The cylinders are circular in cross section, having



% Orthochems & allochems	% Matrix	Silt-sized minerals (% of matrix)			Max. size of silt (µm)	
		Total	Quartz & Feldspar	Clay	Quartz & Feldspar	Clay
4	96	13	1	12	13	18
7	93	18	3	15	25	41
6	94	22	8	14	23	19
12	88	12	1	11	33	36
15	85	8	1	7	16	22
55	45	18	7	11	37	33
22	78	17	1	16	34	35
16	84	14	2	11	13	32
19	81	17	2	14	15	34
49	51	14	1	13	39	41
11	89	17	3	14	23	32
8	92	16	2	14	22	13
35	68	14	2	12	24	15
10	90	19	4	15	33	25
46	54	15	3	12	35	24
5	95	10	3	7	16	38
10	90	14	3	11	17	26
9	81	16	4	14	18	30

Table 1. Petrographic composition of clayshales and limestones.

a diameter of 2 cm and a length of 7 cm. Teeter (1978) described similarly inclined cylinders within Ordovician shales and interpreted them as being touch marks made by cephalopods; they appear to be cephalopod body chambers.

Pyritized fecal pellets, a few centimeters in length, are abundant within a concretionary zone (31.5 meters above the lowest exposure of Ledyard) that is located in the upper portion of the Wanakah Member (Fig. 3). Their sculpting is simple, as they have only an end construction.

Fossils of terrestrial plants are quite common within the Middle Devonian rocks in central and eastern New York but they are rare in the western portion of the state. A few carbonized plant axes were the only examples observed within the Ludlowville Formation at Cazenovia Creek.

Principal Subdivisions

At Cazenovia Creek the Ludlowville Formation can be subdivided into three units based on the distribution and preservation of fossils. My observations are supplemented with data from previous studies in this region (Grabau, 1899; Beerbower and others, 1969; Bray, 1971; McCollum, 1980; Miller, 1982). The subdivisions are: 1) the lower and middle portions of the Ledyard Member, 2) the upper portion of the Ledyard Member and the lower portion of the Wanakah Member (the *Pleurodictyum Zone*), and 3) the middle and upper portions of the Wanakah Member.

The first unit contains only a few shelly fossils, mostly styliolinids, ostracodes, and four genera of brachiopods — *Devonochonetes, Leiorhynchus, Lingula*, and rarely, *Mucrospirifer* (McCollum, 1980). Although these fossils occur throughout this subdivision, their abundance varies considerably from bed to bed; most bedding planes are devoid of calcareous fossils, while some others have dense accumulations of only a single taxon. This unit is further characterized by a large proportion of concretionary beds, most of which contain numerous styliolinid shells. McCollum (1980) has reported that many of the concretionary and the argillaceous-limestone beds also contain molds of nuculid bivalves and that one of the limestone beds (located at 10.8 meters above the lowest exposure of the Ledyard) contains molds of gastropods and cephalopods.

The second unit is very fossiliferous, containing bivalves, brachiopods, bryozoans, corals, crinoids, ostracodes, and styliolinids (McCollum, 1980; Miller, 1982). These fossils are distributed throughout the clayshale as diverse assemblages upon the individual bedding planes. On a few bedding planes, the taphonomy of the skeletal material suggests current action; Miller (1982) reported aligned rugosan corallites and accumulations of brachiopod shells which he interpreted as evidence of storm-wave activity. Trace fossils also are abundant within this unit, and include *Zoophycos* feeding traces (Miller, 1982) and pyritized trail-like traces. Of the four limestone beds that are included (Fig. 3), the lower two are distinctive in that they contain large numbers of cephalopods, bivalves, and gastropods.

The third unit also contains many shelly fossils but they are not as numerous as in the underlying subdivision. Three brachiopods (*Ambocoelia, Athyris,* and *Mucrospirifer*) commonly are preserved in life orientation and, in some instances, are clustered. This clustering is interpreted to have occurred on extensively bioturbated (very fluid) substrates (Beerbower and others, 1969; Bray, 1971). As in the first unit there is an association of a large number of nuculid bivalves with concretionary beds (Bray, 1971).

PETROLOGY

The rocks studied are all fine-grained, consisting primarily of argillaceous micropeloids and minor amounts of silt-sized clay minerals, quartz, pyrite, authigenic calcite, and calcareous and organic-walled fossils (Appendix 1). The abundance of fossils and pyrite varies among stratigraphic subdivisions. The limestones, at least those sampled, are characterized by a micritic calcite cement similar to that contained within the calcium carbonate concretions. The distribution of these constituents is listed in Tables 1, 3, 4, and 5.

Matrix

Matrix comprises the major portion of most thin sections (Table 1). This matrix consists of: 1) micropeloids, and 2) silt-sized grains of monomineralic and monocrystalline illite, chlorite, and quartz.

Micropeloids are the major constituent of every rock sample. They are sub-angular to sub-rounded, and range in diameter from 8 to 17 μm (Fig. 4 and 5). The contacts between adjacent micropeloids are concavo-convex and long (cf. Taylor, 1950), indicating that some compactional deformation has occurred even though no vertical flattenings were observed in the thin sections perpendicular to stratification. No individual minerals were observed within the micropeloids, but they probably consist of clay-sized illite and chlorite. This is because X-ray diffraction of the unfossiliferous shales revealed only illite, chlorite, quartz, pyrite, and sometimes feldspar.

Silt-sized matrix makes up 8 to 22 percent of the rock samples. It consists primarily of clay minerals (7 to 16 percent), but includes some quartz (1 to 8 percent). The maximum size of the silt was determined by measuring the longest diameter of the five largest grains, and for the entire stratigraphic section was found to be medium silt — 24 um for quartz and 29 um for clay minerals (Table 1). Though the percentage of silt and its maximum size varies within this stratigraphic section, no systematic pattern appears either stratigraphically or in association with other characteristics (see Tables 3, 4, and 5).

The silt-sized chlorites, as well as any clay-sized chlorites that might be contained in the micropeloids, are of special interest because this mineral is known to be a source of the ferrous iron in pyrite (Drever, 1971), and pyrite is common in

Distance Between Sample and Concretion	Mg_x^{+2}	Fe_x^{+2}	Al_z^{+3}
0 cm	4.59	0.00	1.41
0 cm	5.38	0.24	0.38
0 cm	4.55	0.04	1.41
0.5 cm	4.55	0.35	1.10
1.0 cm	4.10	0.57	1.33
1.0 cm	4.09	0.89	1.02
2.3 cm	3.87	0.72	1.41
3.0 cm	4.65	0.89	0.46
3.5 cm	4.44	0.07	1.49
3.5 cm	3.70	0.89	1.41
5.0 cm	3.76	0.37	1.87
6.0 cm	4.77	0.13	1.10
6.0 cm	3.62	0.89	1.49
7.5 cm	3.81	0.93	1.26
22.6 cm	5.10	0.28	0.62
24.0 cm	2.79	1.72	1.49
Ave.	4.24	0.56	1.20
S.D.	0.64	0.46	0.41

Table 2. Relative proportions of Mg^{+2}, Fe^{+2}, and Al^{+3}, in octahedral sites of trioctahedral chlorites with composition (Mg^{+2}, Fe^{+2}, Al^{+3}) (Si^{+4}, Al^{+3}) O(OH)g.

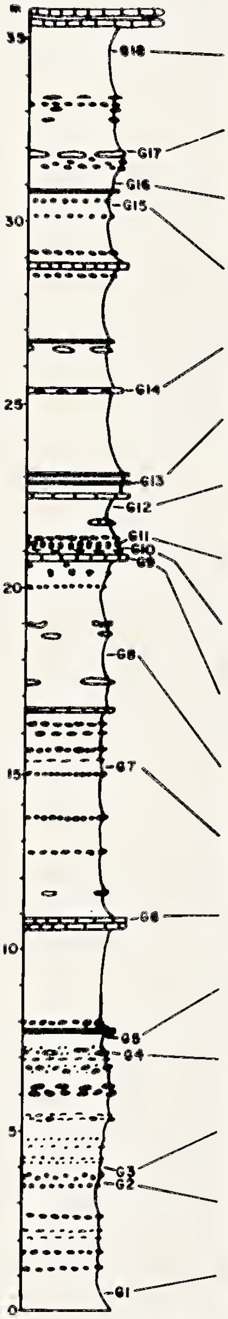
these Ludlowville rocks. Previous studies (Towe, 1961; Liebling and Scherp, 1980) of the chlorites contained within some other Middle Devonian clastic rocks in New York and the probable provenance rocks in New England allow for a comparison of iron content.

Chlorites contained within the rocks from a concretionary zone were analyzed for their ferrous-iron content. The samples consisted of a calcium-carbonate concretion and some nearby clayshales collected from an exposure of Wanakah shale, 31.5 meters above the lowest exposure of the Ledyard Member (Fig. 3). The clayshales within this concretionary zone were collected at various distances from the concretion in anticipation of a possible gradient in ferrous-iron content.

All of the samples from the concretionary zone contain chlorites low in iron and high in magnesium (Table 2; see Appendix 2 for calculations). The chlorites contained within the clayshales have only small proportions of iron but they have a higher iron content than the chlorites contained within the concretion. Overall, only one of the sixteen samples has a Fe^{+2}/Mg^{+2} ratio greater than 0.25; 0.13 is the average ratio.

Skeletal Materials

Most of the shelly fossils have retained their original mineralogy and skeletal structures. However, the mineralogy of crinoid and sponge skeletons has been changed and the aragonite of molluscan shells was dissolved and subsequently not replaced. Many of the crinoid columnals have been extensively micritized, thereby obliterating their stereom morphology. The sponge spicules (monaxon and triaxon) have been replaced by pyrite and, less commonly, by single crystals of calcite. On the basis of taxonomic analogs, the original mineralogy of these spicules was probably opaline silica.



Ostracodes	Trilobites	Crinoids	Fen. Bryozoans	Cten. Bryozoans	Spicules (pyrite)	Spicules (calcite)	Brachiopods	Gastropods	Styliolinids	Tentaculids	Corals	Worm Tubes	Unknown	Total
tr					1	2	tr		tr					3
tr	tr	tr			2		1		1				2	6
					3		tr		tr				2	5
		tr	1		4	1	2						2	10
1	2		tr		2	1	3		tr					9
1	2	1		1	tr	3			25					33
2	6	5	3		1	1	3	tr			tr			21
1			1	tr		1	3		4			2	3	15
tr		1	tr		tr	1	2		10				3	17
	tr		2	9	tr	2	8		16					37
		tr			2				1					3
tr					1		tr		tr					1
						2	2		14	tr				18
					tr				1					1
tr	tr				tr	3	1		37				tr	41
									tr					tr
3		1			tr				tr					4
		tr			2				tr					2

Table 3. Stratigraphic distribution of skeletal material (percent of whole rock).

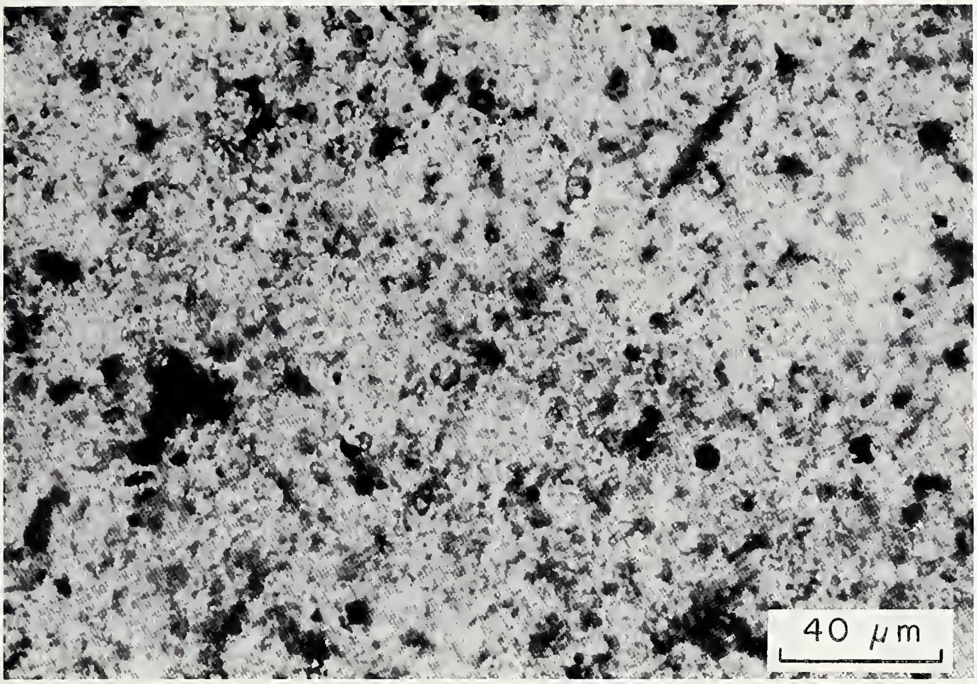


Figure 4. Photomicrograph (plane light) showing the large abundance of micropeloids (G18).

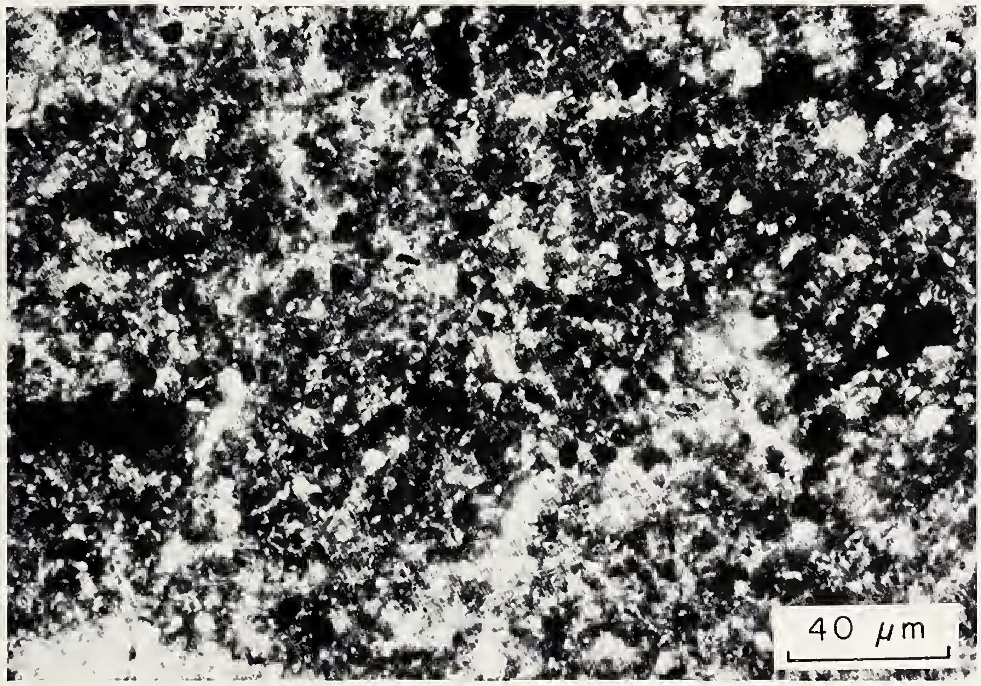
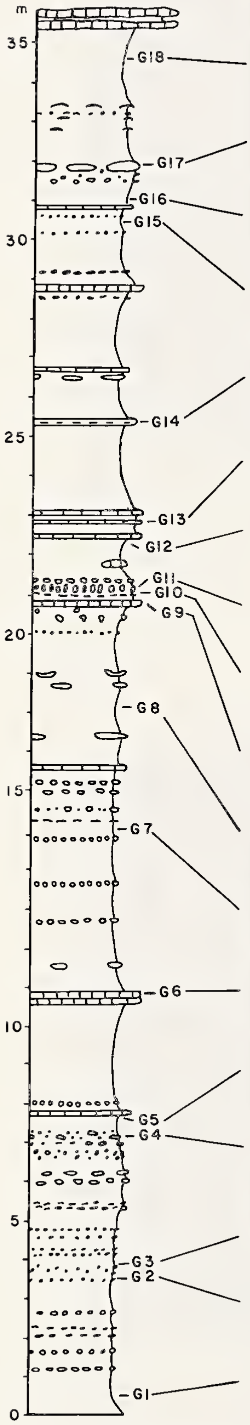


Figure 5. Micropeloids photographed in plane-polorized light (G18).



Disseminated organics	Leiosphaeridiaceae	Chitinozoans	Total: algae & chitinozoans
low	tr	tr	tr
low	tr	tr	tr
low	tr	tr	tr
low	tr	1	1
low	tr	tr	tr
low	tr	tr	tr
low	0	0	0
low	0	0	0
low	0	0	0
low	0	0	0
high	6	0	6
high	4	0	4
high	1	0	1
high	7	0	7
high	1	0	1
high	2	1	3
high	1	2	3
high	2	2	4

Table 4. Stratigraphic distribution of organic-walled tests and disseminated organic material (percent of whole rock).

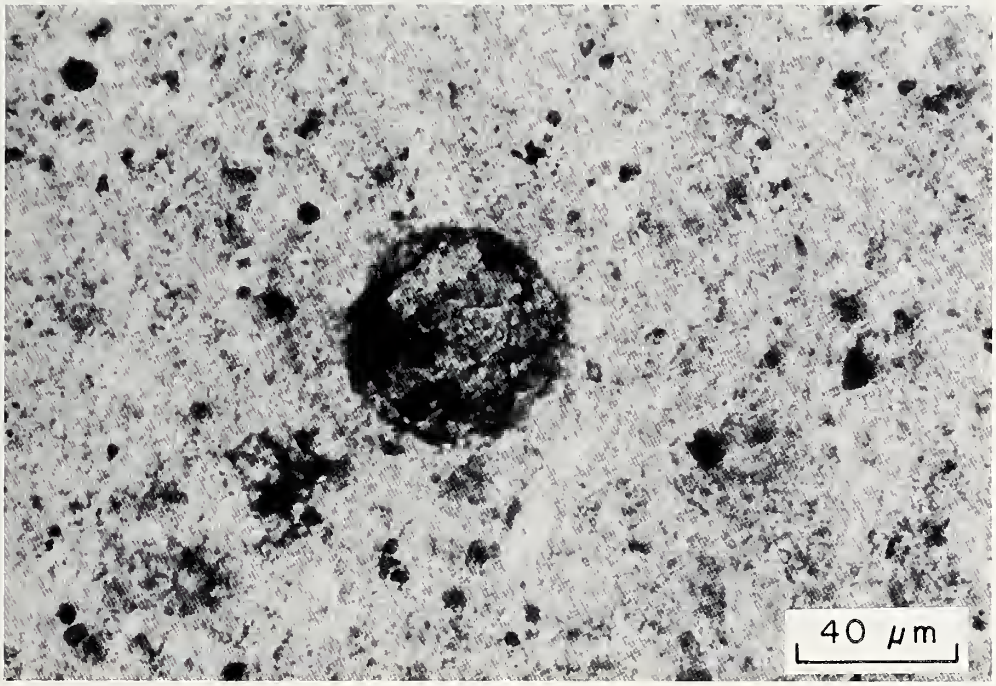


Figure 6. Leiosphaeridiaceae cyst photographed in plane light (G8).

The distribution of shelly fossils in thin sections changes through the stratigraphic section (Table 3). The lower subdivision (lower and middle Ledyard) is not very fossiliferous and those fossils that do occur represent only a few taxa. Styliolinids and sponges are relatively common, while brachiopods, crinoids, and ostracodes are rare. The middle subdivision (upper Ledyard and lower Wanakah) is very fossiliferous. Crinoids, ostracodes, and trilobites are all more numerous here than in the lower subdivision, as are styliolinids and sponges. The uppermost subdivision (middle and upper Wanakah) also is fossiliferous but less so than the middle subdivision. The taxa listed for the middle subdivision are less numerous here, except for sponges, which are most abundant in this upper subdivision.

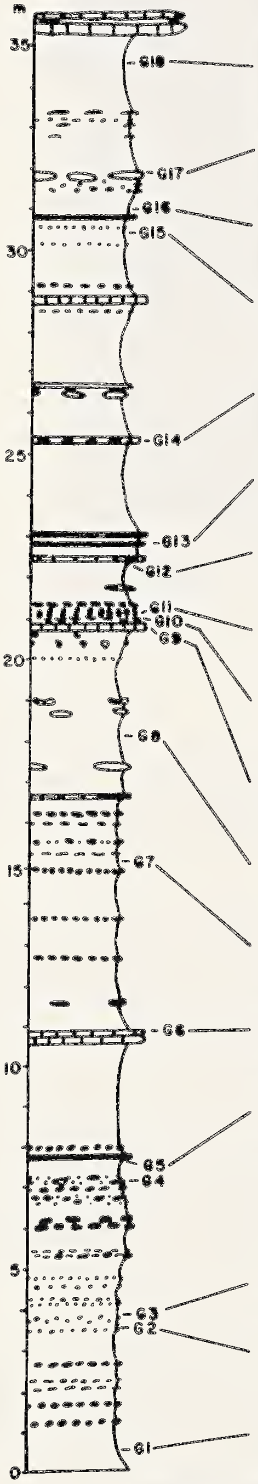
The brachiopod valves within the upper two subdivisions occasionally are articulated, unlike the valves of *Leiorhynchus* and *Devonochonetes* within the lower subdivision. The exteriors of these articulated valves served as substrates for epibionts. Several valves contain borings, most of which are flask-shaped. One pair of *Ambocoelia* valves (G15) also has an algal coating, the only such coating observed in these rocks. A ctenostome bryozoan, *Allonema* (Bassler, 1953, p. G35), was found fossilized as encrustations upon some nuculid bivalves (McCollum, 1980) but was not observed to be associated with brachiopods.

Sand and silt-sized fragments of brachiopod shells are present in the middle subdivision (samples G9 and G14). These are the only occurrences of calcareous sediments that were derived from shell fragmentation.

Organic Material

A large abundance of organic-walled microfossils and disseminated organic material (small unidentifiable fragments) is present within some samples (Table 4). The microfossils are reddish-brown tests that range from 40 to 230 um in diameter. They represent green algae and chitinozoans; no terrestrial-plant spores were observed. Sufficient wall structures are visible to enable taxonomic assignment to family (Leiosphaeridiaceae) for green algae and to genera (*Ancyrochitina*) for chitinozoans (cf. Tappan, 1980). Chitinozoans are tests of uncertain affinity, having been variously considered as fungi, plant-like protists, and metazoan reproductive bodies (Grahn and Afzelius, 1980).

The two varieties of microfossils are easily distinguished. Leiosphaeridiaceae cysts were originally spherical but they can take on an assortment of shapes because of compactional deformation. Most are still spherical (Fig. 6), ranging from 30 to 230 um in diameter. The *Ancyrochitina* specimens observed in the lower subdivision were flask-shaped and are 120 um in longest diameter (Fig. 7).



PYRITE							Micritic calcite
Dissemin-ated	Predepos. pores	Vugs	Replace-ment	Total: primary	Total: secondary	Total	
1	0	tr	1	1	1	2	0
1	0	tr	2	1	2	3	0
1	0	1	3	1	4	5	0
1	0	0	4	1	4	5	0
tr	tr	tr	2	tr	2	2	6
1	0	tr	tr	1	1	2	22
tr	1	2	1	1	3	4	0
1	tr	5	tr	1	5	6	0
1	1	1	tr	2	1	3	0
tr	1	0	tr	1	tr	1	11
2	0	0	2	2	2	4	0
3	0	tr	1	3	1	4	0
tr	tr	0	0	tr	0	tr	13
2	0	0	tr	2	tr	2	0
3	1	tr	tr	4	tr	4	0
2	tr	1	0	2	1	3	0
3	tr	tr	tr	3	tr	3	0
3	tr	tr	2	3	2	5	0

Table 5. Stratigraphic distribution of pyrite and micritic calcite (percent of whole rock).

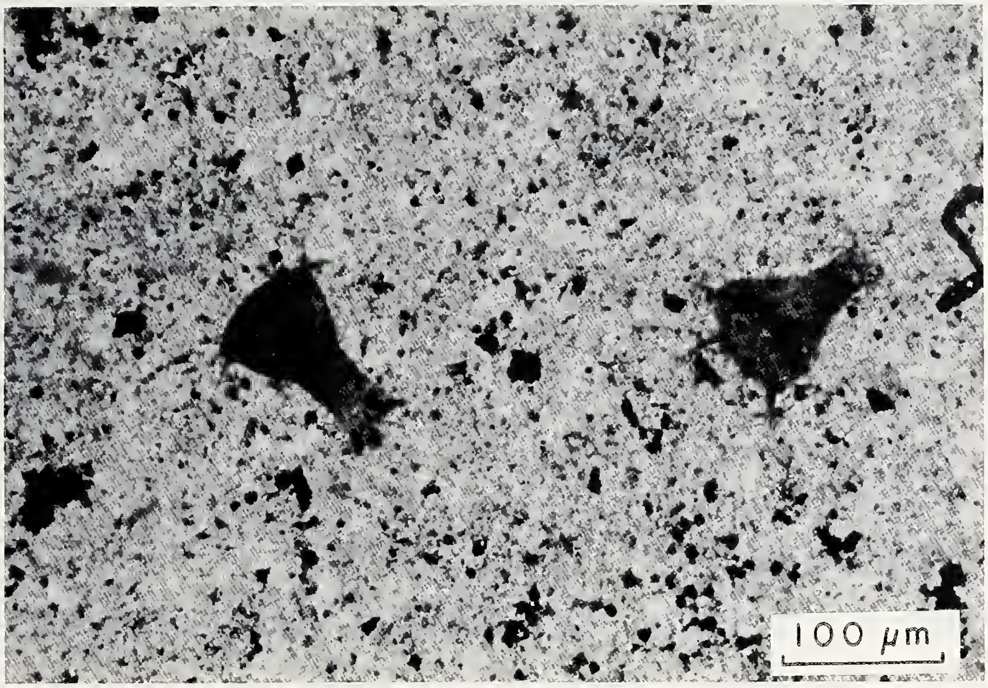


Figure 7. Ancyrochitinid-chitinozoan tests photographed in plane light (G2).

The abundance of algal microfossils and the concentration of disseminated organic material varies among the three stratigraphic subdivisions (Table 4). Both are most abundant in the lowest subdivision (G1-G8), as shown in Figure 8. The middle subdivision (G9-G13) completely lacks observable organics; the uppermost subdivision (G14-G18) contains a few tests but only as fragments (Fig. 9) and some disseminated organic material.

The stratigraphic distribution of chitinozoans is similar to that of the algal microfossils (Table 4). They are abundant only in the lower and middle portions of the Ledyard Member where they are preserved whole. Fragments of chitinozoans, easily distinguished by their reddish tint, occur in trace amounts in the middle and uppermost portions of the Wanakah Member. The similarity in the distribution of chitinozoans and algal cysts suggests that the taphonomy of chitinozoans was similar to that of the algae.

Traces

All of the trace fossils observed in thin section were made by deposit feeders and can be grouped into two general categories. The first category consists of cylindrical patches present within four samples (G1, G4, G5, and G14) and they constitute no more than 5 percent of the rocks. These traces have, for the most part, a horizontal orientation. Their lack of organic particles contrasts sharply with the adjacent matrix where organic-walled microfossils and disseminated organic material are abundant (Fig. 10). These cylinders are interpreted as being feeding traces of a nonsorting, detritus (sediment) feeder. The second category of traces consists of circular features that contain concentrically arranged shell debris. They are approximately 6.3 mm in diameter and exhibit dark-brown rims, 0.8 mm thick. These traces are present within two samples (G9 and G13) and have been identified by Miller (1982) as *Zoophycos* feeding traces.

Pyrite

Pyrite is present in all of the rock samples and has several modes of occurrence (Table 5). These are grouped into two categories based on whether the pyrite fills pores which are primary or secondary (cf. Choquette and Pray, 1970). Primary pores are those within the sediment at the time of, or just after, deposition; they include interparticle and intraparticle pores. Secondary pores are those formed after deposition by the dissolution of unstable grains or cements. The secondary porosity is termed fabric-selective when only isolated grains or minerals have been dissolved. The filling of fabric-selective pores, in this study, is called replacement. Vuggy porosity is non-fabric-selective in that many grains have been dissolved and the resulting pore boundaries are irregular with respect to grain boundaries.

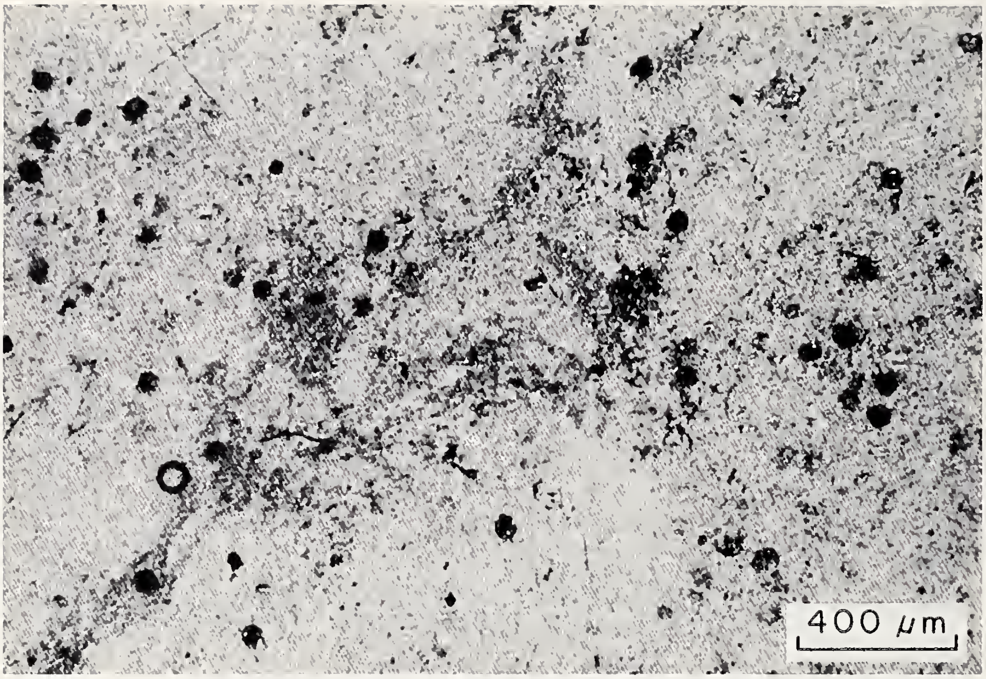


Figure 8. Organic-rich clayshale photographed in plane light. Large spheres are Leiosphaeridiaceae cysts, smaller and opaque grains are pyrite (G5).

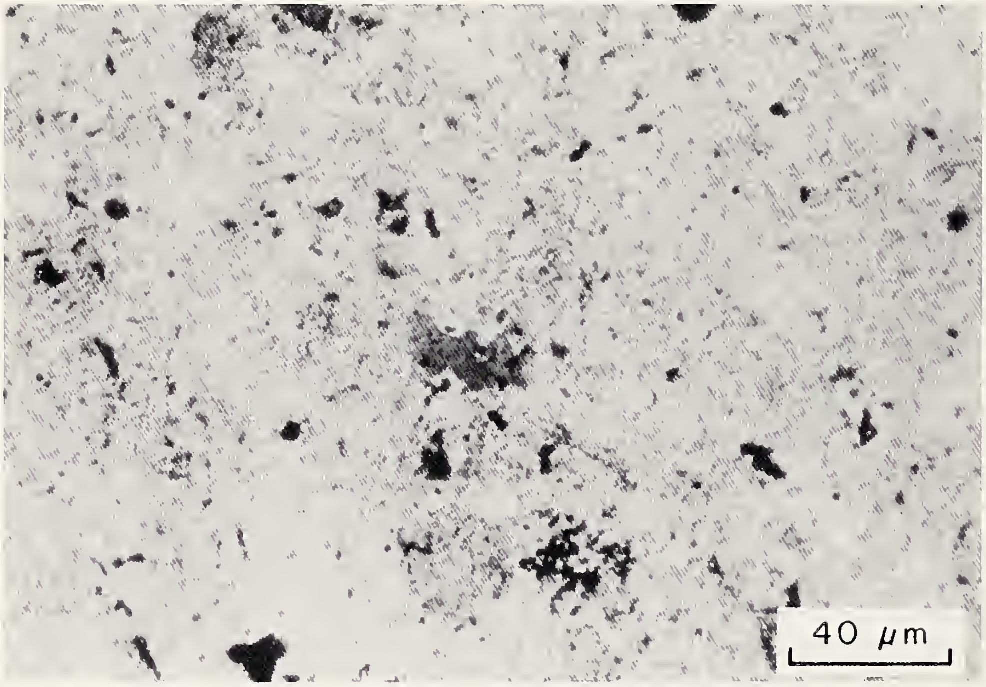


Figure 9. Fragment of a microfossil test photographed in plane light (G18).

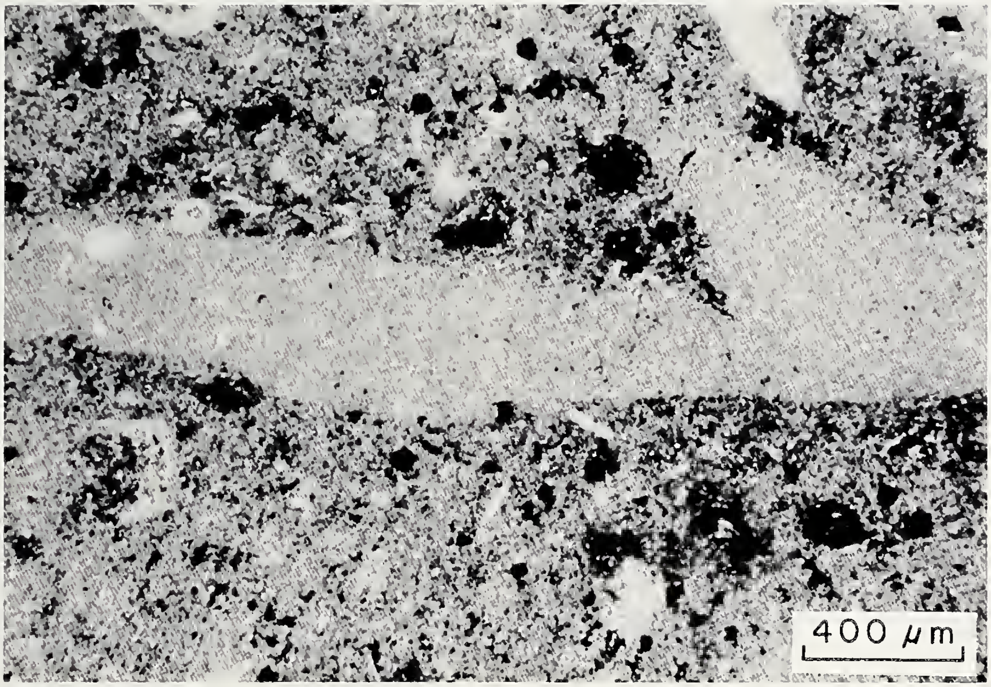


Figure 10. Feeding traces made by a non-sorting, detritus feeder, photographed in plane light (G4).

Disseminated pyrite occurs as spherical aggregates (4 to 140 um in diameter), appears in all of the rock samples, and accounts for much of the pyrite present (Table 5). It probably filled the primary, interparticle pores that were occupied by water. Some disseminated pyrite might have been precipitated in spaces originally occupied by organic particles (rather than in waterfilled voids) but this is unlikely because no organic rims were observed. Disseminated pyrite rarely occurs in the feeding traces (Fig. 10) that have a low content of organic material. This suggests that disseminated pyrite was precipitated near the organic material that was catabolized by sulfate-reducing bacteria.

The remaining primary-pore pyrite is found within the intraparticle pores of calcareous skeletons. This mode of occurrence is only of minor significance in most clayshale samples. However, in three argillaceous limestone samples (G6, G9 and G13) and in the clayshale sample from the *Pleurodictyum* Zone (G12) the intraparticle pores contain much pyrite. The pyrite occupies the following intraparticle pores: arthropod exoskeleton pores, bryozoan zooecia, borings in shells, crinoid stereom, chambers within organic-walled microfossils, internal cavities of styliolinids, and the spaces between articulated valves of brachiopods and ostracodes.

Most of the secondary-pore-filling pyrite occurs in fabric-selective pores that were created by the dissolution of sponge spicules. Framboidal-pyrite aggregates that are shaped as elongate rods, 10 to 20 um wide and several millimeters in length, are interpreted as being monaxon (and sometimes triaxon) sponge spicules that have been pyritized. A few of the spicules were not pyritized but instead were calcified (see Table 3). The monaxon and triaxon spicules within modern sponges are composed of either aragonite or opal. If these spicules originally were aragonite, then their large surface area-to-volume ratio would have resulted in dissolution more rapid than that of the aragonitic molluscan shells; the spicules could have dissolved while in the sulfidic, diagenetic environment, unlike the molluscan shells that usually dissolved in a post-sulfidic environment. Yet opaline spicules could also have been dissolved in the sulfidic, diagenetic environment. An elevated pH of the pore waters would not only enhance the dissolution of opal but would also allow calcite to replace the opal (Blatt and others, 1980, p. 348). The pH could have been increased by a buildup of ammonia. Ammonia is produced by the anaerobic decomposition of proteinaceous compounds and its concentration increases during shallow burial (Berner, 1980, p. 138).

Non-fabric selective pores are filled with pyrite and occasionally calcite but they occur only sparsely. Pyrite-filled vuggy pores are present in about one-half of the rock samples but are neither abundant nor large. A second, even rarer type of secondary pore was spherical and presently is filled with pyrite and calcite (Wygant, 1983, p. 51). These bodies are present in trace amounts within only three samples (G11, G15, and G18). They average about 240 um in diameter and contain numerous pyrite spheres (30 um in diameter) that, along the periphery, protrude into the micropeloidal matrix. The calcite contains no definite skeletal structures which, together with its association with pyrite, suggest that it is authigenic and fills secondary pores that were spherical. These spherical pores could have been formed by three processes: the dissolution of siliceous organisms, the complete decomposition of organic-walled microfossils, or the formation of gas (methane?)vacuoles.

The occurrence and abundance of pyrite exhibit both lithologic and stratigraphic correlatives (Table 5). The limestone samples have less total pyrite (1.0 percent on average) than the clayshale samples (3.7 percent on average). Although the total amount of pyrite in the clayshales of the Ledyard and Wanakah members is almost equal (3.6 and 3.8 percent respectively), primary-pore-filling pyrite is most abundant in the Ledyard. This larger amount of primary-pore-filling pyrite corresponds to a higher content of organic material.

Calcite Cement

Authigenic calcite occurs as micrite and spar. The distinction between these two calcites is more than that of crystal size; each occurs in different types of pores. Micrite is restricted to interparticle pores, whereas sparry calcite is restricted, for the most part, to intraparticle pores. Micritic calcite is present and abundant in only four rock samples (Table 5). In the field these four rocks weather as limestones. Sparry calcite is present in many samples but this authigenic calcite is not tallied separately from the fossils because it fills skeletal cavities. No sparry calcite occurs disseminated within the micropeloidal matrix but minor amounts do occur as overgrown on the exterior of some shells.

The sparry calcite comprises a significant volume of a few rock samples. Bladed crystals syntaxially fill, partly to completely, some styliolinid cavities and the zooecia of the few ctenostome bryozoans. These calcite filled styliolinid shells are most numerous in three limestone samples (G6, G9, and G13) and in one clayshale sample (G4). Equant crystals that are medium crystalline (10 to 320 um) occur between the articulated valves of brachiopods and ostracodes and rarely within the algal tests. Fibrous calcite occurs mostly as single crystals (10 to 20 um wide and several millimeters in length) replacing the opal or aragonite in sponge spicules. Minor amounts of fibrous calcite are present as very finely crystalline overgrowths on the exterior of a few styliolinid shells (samples G9 and G10).

Lithification Features

Porosity reduction within the Ludlowville clayshales and argillaceous limestones was extensive because argillaceous sediments lose between 60 and 80 percent of their porosity during burial (Weller, 1959). The depositional porosity probably was interparticle for the most part and was destroyed as the result of compaction and cementation. Compactional processes were of two types: mechanical compaction and pressure solution (cf. Meyers, 1980). The former caused the slight deformation of micropeloids and the fracturing of calcareous skeletons, while the latter caused the interpenetration of calcareous skeletons. Cementational processes resulted in the precipitation of iron sulfides and micritic calcite within interparticle pores.

Mechanical compaction destroyed the uncemented, interparticle pores. These interparticle pores were occluded during burial by the slight deformation of the micropeloids. In a few samples, mechanical compaction fractured the calcareous skeletal material. This fracturing is restricted to the five most fossiliferous samples: three argillaceous limestones (G6, G9, and G13) and two clayshales (G4 and G12). The skeletal material in four of these samples consists primarily of styliolinid shells and it is these shells that are often fractured. In the remaining sample (G12), it is the articulated ostracode shells and trilobite segments that are fractured.

Only some of the Ludlowville sediments were cemented with micritic calcite; these are now argillaceous limestones and concretions. A relative dating of micritic calcite is difficult because it does not occur within intraparticle pores. This absence from intraparticle pores might suggest that micritic calcite was precipitated later than pyrite and sparry calcite (i.e., the shells were already filled). Micritic calcite certainly was precipitated after the onset of pyrite precipitation because pyrite is encased within the limestones (as reported here) and the concretions (Jordan, 1968). The abundance of micritic calcite might be a reflection of the sediment's porosity at the time of precipitation. But this would require that the micritic calcite completely filled all of the available pores and that no displacive crystal growth occurred.

Sparry calcite and pyrite, though important indicators of chemical conditions during diagenesis, were only of minor importance in the lithification of these sediments. These two cements can be ordered sequentially on the basis of their textural relationships within intraparticle pores. The infillings of styliolinid shell cavities reveal that sparry calcite was precipitated during the precipitation of pyrite. Pyrite occasionally occupies the apex of the cavities, occurs as inclusions within the cavity-filling sparry calcite, and fills the outer portions of the shell cavities. This textural relationship indicates that the sparry calcite was precipitated in a sulfidic, diagenetic environment.

Pressure solution was also of only minor importance in the lithification of these rocks. Solution-shell contacts are present only in the same five samples (G4, G6, G9, G12 and G13) that contain fractured shells. In sample G12, trilobite fragments commonly interpenetrate brachiopod shells and crinoid columnals. In the other four samples, solution-shell contacts are common between styliolinid shells. However, only in the sample with the most styliolinid shells (G4) was there any substantial loss of shell material. A relative dating of pressure solution is difficult because solution-shell contacts were not observed cross-cutting any sparry calcite or pyrite. This absence of cross-cutting relationships might imply that pressure solution was contemporaneous with the precipitation of pyrite. If so, then pressure solution of styliolinid shells occurred in a sulfidic, diagenetic environment.

INTERPRETIVE SUMMARY

The biological features within these Middle Devonian rocks can be used to reconstruct the sedimentary environments. This is because the biological activities upon and within the substrates were effectively controlled by physical and chemical variables during deposition and early diagenesis.

The concentration of dissolved oxygen within the seawater controlled the presence of benthic animals during the deposition of the Ledyard and Wanakah. The benthic faunas of low diversity could have been caused by several conditions: brackish seawater, hypersaline seawater, or oxygen-poor seawater. The stratigraphic equivalence of these gray shales to black shales (in central New York) initially suggests that it was low concentrations of dissolved oxygen that occasionally excluded the benthic animals from the substrates. It is unlikely that the seawater was ever brackish because brackish seawater would not have been able to supply enough sulfate to form the large abundance of pyrite that occurs throughout the Ledyard and Wanakah. Hypersaline seawater also is unlikely because none of the minerals that characterize evaporite sedimentation is present.

The lower and middle portions of the Ledyard represent deposition upon a substrate immediately overlain by seawater, containing little or no oxygen. This stratigraphic interval contains only a few varieties of shelly fossils, mainly ostracodes, styliolinids, and disarticulated valves of the brachiopods *Leiorhynchus* and *Devonochonetes*. The ostracodes (if pelagic) and the styliolinids (Tucker and Kendall, 1973) probably lived in the well-oxygenated seawater closer to the sea's surface. *Leiorhynchus* was a benthic animal, and *Devonochonetes* must have been epibenthic because it lacked a functional pedicle. So, the occurrences of *Devonochonetes*, *Leiorhynchus*, and ostracodes (if benthic or epibenthic) on some bedding surfaces within the lower and middle Ledyard probably represents brief, oxygenated conditions immediately above the substrate. The feeding traces that were made by nonsorting, deposit-feeding animals on some bedding surfaces within this stratigraphic interval also represent brief oxygenated conditions. These traces, when present, occupy only a small portion (less than 5 percent) of a bedding surface, indicating that the concentration of dissolved oxygen was low and that the low-oxygen conditions did not last long enough for the deposit-feeding animals to consume much of the organic-rich sediment. Consequently, the rocks of the lower and middle Ledyard contain an abundance of organic material.

The amount of organic material within the Ledyard and Wanakah was controlled by the extent of benthic animal feeding. A reciprocal relationship exists between the distribution of fossilized benthic animals (Fig. 3 and Table 3) and that of organic-walled microfossils (Table 4). Where fossils of benthic animals are rare (the lower and middle Ledyard), the microfossils are abundant and intact. But where fossils of benthic animals are abundant (the upper Ledyard and all of the Wanakah), the microfossils are rare and, when present, usually fragmented. The disseminated organic material displays a stratigraphic distribution similar to that of the organic-walled microfossils.

The upper portion of the Ledyard and the entire Wanakah represent deposition upon a well-oxygenated substrate. These rocks contain the fossils of many benthic animals: brachiopods, bryozoans, corals, crinoids, and sponges. Some of these benthic animals (rugose corals and the brachiopods *Ambocoelia*, *Athyris*, and *Mucrospirifer*) are occasionally preserved in their life position. But much of the skeletal material in the lower Wanakah (Miller, 1982) and in the uppermost Wanakah (Beerbower, pers. comm.) was apparently reworked by storm waves. None of the rocks within the Ledyard and Wanakah Members display any features that are indicative of sedimentation within normal wave base.

Fragmentation of microfossil tests during the formation of the middle and upper Wanakah was caused by either the storm waves or by the mechanical digestive processes of detritus-feeding animals. Such reworking of sediment by sedimentary processes is believed to have fragmented the organic-walled microfossils contained within some fine-grained Silurian rocks (Smith and Saunders, 1970). However, the mechanical digestive-processes of animals could also have caused test fragmentation in the Wanakah. The type of benthic animal (deposit feeder or suspension feeder) that might have been responsible is unclear. But the deposit feeding animals that left the feeding traces within the lower and middle Ledyard were not involved because their traces lack test fragments.

The micropeloids are not diagenetic features that were produced by burial metamorphism. None of the diagnostic characteristics of authigenic clay minerals (e.g., convolute edges, stacked platelets, and grain overgrowths) are present (Scholle, 1979). Furthermore, the diagenetic alternation of clay minerals occurs at temperatures higher than the 60°C temperature (Legall and others, 1981) that the Hamilton Group in western New York reached during burial. A diagenetic origin for the clays also is discounted by the abrupt contact between the feeding traces and the adjacent matrix (Fig. 10). These contacts probably would have been obliterated had the clay minerals been recrystallized.

Pellet production by suspension-feeding animals was the major process by which the sediments of the Ledyard and Wanakah were deposited. The micropeloids were produced either by pelagic, suspension-feeding animals or by soft-bodied, benthic, suspension-feeding animals. Shelly benthic animals could not have produced the micropeloids because a portion of the depositional sequence (the lower and middle Ledyard) lacks such fossils but still contains an abundance of micropeloids.

Biotic processes continued to be important even after dissolved oxygen was depleted from the pore waters. Sulfate-reducing bacteria (obligate anaerobes) catabolized organic material and produced hydrogen sulfide, which then reacted with the iron from clay minerals and iron oxides to form iron sulfides. The iron sulfides and the less common micritic calcites are the only cements that filled interparticle pores. Even though iron sulfides were only of minor importance in the lithification of these clayshales and limestones, their presence has important geochemical implications.

The amount of iron sulfide precipitated within primary pores was dependent upon the abundance of organic material within the sediments when the pore water became anoxic. The pore water within the lower and middle Ledyard became anoxic early during burial and perhaps was even anoxic at the sediment-water interface. As a result, much labile organic material was available to the sulfate-reducing bacteria and iron sulfides subsequently were produced. Further support of the relationship between organic material and primary-pore pyrite is provided by the feeding traces within portions of the lower and middle Ledyard. Unlike the adjacent sediment, the feeding traces lack organic material and, correspondingly, contain no primary-pore pyrite. Also indicated by these feeding traces, the amount of primary porosity and the concentration of dissolved sulfate did not control the abundance of iron sulfides. The traces have, at the very least, the same primary porosity and the same concentration of dissolved sulfate as that of the adjacent organic-rich sediment.

Sparry calcite was precipitated within a sulfide diagenetic environment, apparently as the result of an increase in the concentration of carbonate ion. The environment of precipitation is revealed by the occurrence of sparry calcite between two generations of pyrite (within some styliolinid-shell cavities). In anoxic-and-sulfidic sediments, a variety of reactions can often lead to an increase in carbonate alkalinity. For instance, carbonate ions can be generated by bacterial sulfate reduction and by the anaerobic decomposition of proteinaceous compounds (Berner, 1969; 1980, p.129 – 130).

The low proportions of iron within the chlorites are not related to provenance, instead, they represent a post-depositional change. Various workers have reported that the chlorites occurring within Middle Devonian near-shore rocks and rocks in their provenance have a high content of iron, containing at least as much iron as magnesium (Towe, 1961; Liebling and Scherp, 1980). In the Ledyard and Wanakah sediments, iron released from these detrital chlorites reacted with aqueous hydrogen sulfide and formed pyrite. Magnesium, the most abundant cation in seawater, subsequently filled those octahedral sites vacated by iron. The chlorites that became encased within the concretions released the most iron, perhaps because the reducing conditions were greatest at these locations.

The availability of sedimentary iron probably determined the amount of pyrite that was precipitated. Three lines of evidence suggest this: 1) the total amount of pyrite within the clayshales of the Ledyard (3.6 percent) and the Wanakah (3.8 percent) is almost equal even though they have different amounts of organic material; 2) the clayshales have four times more pyrite than do the limestones, possibly because the clayshales contained more clay minerals (a major source of iron) and have undergone more compactional enrichment; and 3) the detrital chlorites have been depleted of their exchangeable iron.

Dissolution of aragonitic molluscan shells resulted in the formation of argillaceous limestones and concretions. Large assemblages of nuculid bivalves, gastropods, and cephalopods are present as molds within the micritic-cemented lithologies (McCollum, 1980). The dissolution of these aragonitic shells contributed carbonate ions to the pore waters and this led to the precipitation of micritic calcite within interparticle pores.

Carbonate ions that were released by the dissolution of molluscan aragonite were transported laterally by pore water. This is supported by three lines of evidence: 1) most concretions are laterally elongated; 2) the elongation axes of concretions within some beds are parallel in direction; and 3) micritic calcite is, in general, restricted to the beds from which the aragonite had been dissolved. Vertical expulsion of this carbonated pore water was unusual and is indicated by only a single feature – vertically elongated concretions located at 20.9 meters above the lowest exposure of the Ledyard. These vertically elongated concretions immediately overlie an argillaceous-limestone bed, suggesting that the source of carbonate ions was this underlying limestone.

Most of the depositional interparticle porosity was destroyed during burial by compaction. The overburden pressure caused the micropeloids to become slightly deformed in shape. This led to a tighter packing of the micropeloids, thereby occluding much of the intraparticle porosity.

The micropeloids must have been somewhat resistant to mechanical compaction because they remain as individual entities and are not vertically flattened. It is most likely that they were originally dense from intense mechanical compaction within the digestive system of the suspension-feeding animals that produced them. The only other possibility is that the micropeloids were internally cemented with aragonite shortly after they were produced. However, this is unlikely because aragonite was dissolved from the sediments of the Ledyard and Wanakah very early during burial; any aragonite that might have been present within the micropeloids would have been dissolved before overburden pressure became significant.

CONCLUSIONS

Fluctuations in the abundance of animals living upon and within the Ledyard and Wanakah sediments during deposition and early diagenesis resulted in petrologic variations. The fluctuations in biota were themselves a consequence of variations in the concentration of dissolved oxygen at the sediment-water interface. The lower and middle portions of the Ledyard Member contain relatively few fossils of benthic animals because these sediments were poorly oxygenated. The low amount of aerobic activity allowed for the preservation of organic-walled microfossils and disseminated organic material. Some of this organic material subsequently was utilized by anaerobic, sulfate-reducing bacteria, resulting in the formation of much primary pore-filling pyrite. In contrast, the sediments of the upper Ledyard and the Wanakah represent deposition upon a well-oxygenated substrate. Fossils of benthic animals and burrows are abundant here and organic material and primary pore-filling pyrite are sparse.

The accumulation of aragonitic molluscan shells and their subsequent dissolution resulted in limestones and concretions. The complete dissolution of these aragonitic shells during early diagenesis supplied carbonate ions for the precipitation of micritic-calcite cement. The pore water in these beds flowed laterally, thereby restricting the micritic calcite to the beds containing aragonitic shells.

Most of the sediments in these two members were deposited as pellets formed by suspension-feeding animals, either pelagic or soft-bodied epibenthic. The slight deformation of these micropeloids (aggregates mostly of clay) was the major mechanism by which the porosity of the sediments was reduced during burial.

ACKNOWLEDGMENTS

This paper is based on my master's thesis at the State University of New York at Binghamton. Reviews by Carlton Brett, Merrell Miller, Gordon Wood, and Donald Woodrow uncovered several errors and made the manuscript more readable.

Several others assisted in this research project. Dave Tuttle (SUNY-Binghamton) photographed the drawings; Matt Gubitosa produced the photographic prints. The field study was partially supported by a Grant-In-Aid from Sigma Xi. Word processing services were provided by Edwards and Kelcey, Inc.

REFERENCES CITED

Aller, R.C. 1977. The influence of macrobenthos on chemical diagenesis of marine sediments. Doctoral dissertation, Yale University, 600 p.

Baird, G.C. 1979. Sedimentary relationships of Portland Point and related Middle Devonian rocks in central and western New York. New York State Mus. Bull. 433, 24 p.

_________ 1981. Submarine erosion on a gentle paleoslope: a study of two discontinuities in the New York Devonian. Lethaia 14:105-122.

_________ and Brett, C.E. 1981. Submarine discontinuities and sedimentary condensation in the Upper Hamilton Group (Middle Devonian): examination of marine shelf and paleoslope deposits in the Cayuga Valley. New York State Geol. Assn. Guidebook, 53rd Ann. Mtg., SUNY at Binghamton, p. 115-145.

_________ and Brett, C.E. 1983. Middle Devonian detrital shelf to-basin ramp: Ludlowville lithofacies and faunal associations in west-central New York State. Geol. Soc. Amer. Abstr. with Programs 15(3):170.

Bassler, R.S. 1953. Ctenostomata. *In* Moore, R.C., ed., Treatise on invertebrate paleontology, Part G, Bryozoa. Geol. Soc. Amer. and Univ. of Kansas Press, Lawrence, p. G35.

Beerbower, J.R. 1981. A paleomicroecologic study of deposition and early diagenesis in a mid-Devonian marine shale. Geol. Soc. Amer. Abstr. with Programs 13:406.

_________ and others. 1969. A "'paleomicroecologic" study of Mid-Devonian marine assemblages (Hamilton Group) in western New York. North Amer. Paleont. Conv. Field Trip #5, p. 48-78.

Berner, R.A. 1969. Chemical changes affecting dissolved calcium during the bacterial decomposition of fish and clams in sea water. Marine Geology 7:253-274.

_________ 1970. Sedimentary pyrite formation. Amer. J. Sci. 268:1-23.

_________ 1980. Early diagenesis: A theoretical approach. Princeton Univ. Press, Princeton, 241 p.

_________ 1981. A new geochemical classification of sedimentary environments. J. Sediment. Petrol. 51:359-365.

Blatt, H., Middleton, G., and Murray, R. 1980. Origin of sedimentary rocks. Prentice-Hall, Englewood, New Jersey, p. 348.

Bray, R.G. 1971. Ecology and "life history" of Mid-Devonian brachiopod clusters, Erie County, New York. Doctoral dissertation, McMaster University, 162 p.

Brett, C.E. and others. 1983. Paleoenvironmental models of the Hamilton Group: cyclic and episodic deposition in a stratified basin. Geol. Soc. Amer. Abstr. with Programs 15(3):172.

Brindley, G.W. and Gillery, F.H. 1956. X-ray identification of chlorite species. Amer. Mineral. 41:169-186.

Broadhead, R.F. and others. 1982. Stratigraphic and sedimentologic controls of gas in shale — example from Upper Devonian of northern Ohio. Amer. Assn. Petroleum Geol. Bull. 66:10-27.

Byers, C.W. 1974. Shale fissility: relation to bioturbation. Sedimentology 21:479-484.

Carney, R.S. 1982. Bioturbation and biodeposition. *In* Boucot, A.J., ed., Marine Paleoecology. Academic Press, New York, p. 357-399.

Choquette, P.W. and Pray, L.C. 1970. Geologic nomenclature and classification of porosity in sedimentary carbonate rocks. Amer. Assn. Petroleum Geol. Bull. 54:207-250.

Cooper, G.A. 1930. Stratigraphy of the Hamilton Group of New York. Amer. J. Sci. 219:116-134, 214-236.

Curtis, C.D. 1980. Diagenetic alteration in black shales. J. Geol. Soc. London 137:189-194.

Dick, V.B. and Brett, C.E. 1982. Taphonomy and depositional environment of Middle Devonian (Hamilton Group) pyritic fossil horizons in western New York. Geol. Soc. Amer. Abstr. with Programs 14:15.

Drever, J.I. 1971. Magnesium-iron replacement in clay minerals in anoxic marine sediments. Science 172:1334-1336.

Edzwarld, J.K. and O'Melia, C.R. 1975. Clay distributions in recent estuarine sediments. Clays and Clay Minerals 23:39-44.

Folk, R.L. 1968. Petrology of sedimentary rocks. Hemphill's, Austin, p. 37.

Friedman, G.M. and Johnson, K.G. 1966. The Devonian Catskill deltaic complex of New York, type example of a "tectonic delta complex." *In* Shirley, M.L. and Ragsdale, J.A., eds., Deltas in their Geologic Framework. Houston Geol. Soc., p. 171-188.

Gibbs, R.J. 1977. Clay mineral segregation in the marine environment. J. Sediment. Petrol. 47:237-243.

Grabau, A. 1899. Geology and paleontology of Eighteen Mile Creek and the lake shore sections of Erie County, New York. Buffalo Soc. Nat. Sci. Bull. 6, 403 p.

Grahn, Y. and Afzelius, B.A. 1980. Ultra-structural studies of some chitinozoan vesicles. Lethaia 13:119-126.

Gutjahr, C.C.M. 1966. Carbonization measurements of pollen grains and spores and their application. Leidse Geologische Mededelingen 38:1-29.

Hall, J. 1857. Descriptions of Palaeozoic fossils. New York State Cab. Nat. Hist., 10th Ann. Rep. p. 41-186.

Heath, G.R. and others. 1977. Organic carbon in deep sea sediments. *In* Anderson, N.R. and Malahoff, A., eds., The Fate of Fossil Fuel CO_2 in the Oceans. Plenum Press, New York, p. 605-625.

Hower, J. and others. 1976. Mechanism of burial metamorphism of argillaceous sediment: 1. Mineralogical and chemical evidence. Geol. Soc. Amer. Bull. 87:725-737.

Hudson, J.D. 1982. Pyrite in ammonite-bearing shales from the Jurassic of England and Germany. Sedimentology 29:639-667.

Hunt, J.M. 1979. Petroleum Geochemistry and Geology. W.H. Freeman and Company, San Francisco, p. 166-176.

Jordan, F.W. 1968. Genesis of carbonate concretions in the Upper Ludlowville, Middle Devonian of Erie County, New York. Master's thesis, McMaster University, 67 p.

Legall, F.D. and others. 1981. Thermal maturation, burial history and hotspot development, Palaeozoic strata of southern Ontario and Quebec, from conodont and acritarch colour alteration studies. Bull. Can. Petroleum Geol. 29:492-539.

Leibling, R.S. and Scherp, H.S. 1980. Chlorite and mica as indicators of provenance. Clays and Clay Minerals 28:230-232.

Lewis, T.L. 1976. Late Devonian and early Mississippian paleoenvironments, northern Ohio. Geol. Soc. Amer. Abstr. with Programs 8(2):220.

Lundegard, P.D. and Samuels, N.D. 1980. Field classification of fine-grained sedimentary rocks. J. Sediment. Petrol. 50:781-786.

McCave, I.N. 1967. A stratigraphical and sedimentological analysis of a portion of the Hamilton Group (Middle Devonian) of New York State. Doctoral dissertation, Brown University, 381 p.

McCollum, L. 1980. Paleoecology of the Ledyard Shale Member (Middle Devonian: Givetian) of New York. Doctoral dissertation, SUNY at Binghamton, 143 p.

McIver, N.L. 1970. Appalachian turbidites. *In* Fisher, G.W. and others, eds., Studies of Appalachian Geology — Central and Southern. Interscience Publishers New York, p. 69-81.

Meyers, W.S. 1980. Compaction in Mississippian skeletal limestones, southwest New Mexico. J. Sediment. Petrol. 50:457-474.

Miller, K.B. 1982. Paleomicroecologic analysis of depositional environments and sequences, "*Pleurodictyum* Zone," Ludlowville Formation, Middle Devonian of western New York. Master's thesis, SUNY at Binghamton, 131 p.

__________ 1983. Paleomicroecologic analysis of depositional environments and sequences, "*Pleurodictyum* Zone," Ludlowville Formation, Middle Devonian of western New York. Geol. Soc. Amer. Abstr. with Programs 15(3):171.

Pryor, W.A. 1975. Biogenic sedimentation and alteration of argillaceous sediments in shallow marine environments. Geol. Soc. Amer. Bull. 86:1244-1254.

Raiswell, R. 1982. Pyrite texture, isotope composition and the availability of iron. Amer. J. Science 282:1244-1263.

Rhoads, D.C. 1974. Organism-sediment relations on the muddy sea floor. Oceanography and marine biology: An annual review 12:263-300.

Rhoads, D.C. and Young, D.K. 1970. The influence of deposit-feeding organisms on sediment stability and community trophic structure. J. Mar. Res. 28:150-178.

Rickard, L.V. 1975. Correlation of the Silurian and Devonian rocks of New York State. New York State Mus. Map and Chart Ser. 24, 16 p.

Scholle, P.A. 1979. Constituents, Textures, Cements, and Porosities of Sandstones and Associated Rocks. Amer. Assn. Petroleum Geol., Tulsa, 201 p.

Schrader, J.H. 1971. Fecal pellets: role in sedimentation of pelagic diatoms. Science 174:55-57.

Simpson, S. 1975. Classification of trace fossils. *In* Frey, R.W., ed., The Study of Trace Fossils. Springer-Verlag, New York, p. 39-54.

Smayda, T.J. 1971. Normal and accelerated sinking of phytoplankton in the sea. Marine Geology 11:105-122.

Smith, N.D. and Saunders, R.S. 1970. Paleoenvironments and their control of acritarch distribution: Silurian of east-central Pennsylvania. J. Sediment. Petrol. 40:324-333.

Speyer, S.E. and Brett, C.E. 1983. Storm generated "event horizons" in the Middle Devonian Hamilton Group of western

New York. Geol. Soc. Amer. Abstr. with Programs 15(3):171.

Syvitski, J.P.M. and Lewis, A.G. 1980. Sediment ingestion by *Tigripus californicus* and other zooplankton: mineral transformation and other sedimentological considerations. J. Sediment. Petrol. 50:869-880.

Tappan, H. 1980. Paleobiology of plant protists. W.H. Freeman and Company, San Francisco, 1028 p.

Taylor, J.M. 1950. Pore-space reduction in sandstone. Amer. Assn. Petroleum Geol. Bull. 34:701-716.

Teeter, J.W. 1978. Cephalopod touch marks from the Craigleith Member, Whitby Formation (Ordovician) of central Ontario. J. Paleont. 52:178-181.

Towe, K. 1961. Lateral variations in clay mineralogy across major facies boundaries in the Middle Devonian (Ludlowville), New York. Doctoral dissertation, University of Illinois at Urbana, 75 p.

Tucker, M.E. and Kendall, A.C. 1973. The diagenesis and low-grade metamorphism of Devonian styliolinid-rich pelagic carbonates from West Germany: possible analogs of recent pteropod oozes. J. Sediment. Petrol. 43:672-687.

Verwey, J. 1952. On the ecology of distribution of cockel and mussel in the Dutch Waddensee, their role in sedimentation and the source of their food supply. Archives Neerlandaises De Zoologie 10:171-240.

Weller, J.M. 1959. Compaction of sediments. Amer. Assn. Petroleum Geol. Bull. 43:273-310.

Whitehouse, V.G. and others. 1960. Differential settling tendencies of clay minerals in saline waters. Seventh Nat. Conf. Clays and Clay Minerals, p. 1-79.

Wilson, J.L. 1975. Carbonate Facies in Geologic History. Springer-Verlag, New York, 471 p.

Wygant, G.T. 1983. Petrology and paleoecology of the Ledyard and Wanakah Shale Members of Ludlowville Formation, Cazenovia Creek, western New York. Master's thesis, SUNY at Binghamton, 75 p.

APPENDIX 1

METHODS AND MATERIALS

Field Studies

Figure 3 represents a composite section of the Ledyard and Wanakah strata exposed in the bed and banks of Cazenovia Creek between Leydecker and Northrup Roads (see Fig. 1). The stratigraphic locations of argillaceous limestone and concretionary beds and the distribution of fossils also are shown on Figure 3. Most of the lithologic features are reported here for the first time. The distribution of fossils is described in more detail by Grabau (1899), McCollum (1980), and Miller (1982). The exact thickness and lithology of the lowest meter, or so, of the Ledyard Member is indeterminate because of a gap in exposure downstream (and downsection) from Leydecker Road to the outcrop of Centerfield Limestone. Although section was measured only at Cazenovia Creek, the lithologic features observed there were found in the same stratigraphic sequence at two other localities: Buffalo Creek and the shore of Lake Erie.

Eighteen bedding-unit samples were collected from the bed of Cazenovia Creek (Fig. 3). Every sample consisted of seven subsamples, each approximately one square decimeter in area and 2 to 5 mm thick. Several centimeters of intact clayshale were removed from above the bedding surface to be sampled so that the effects of weathering, particularly the oxidation of pyrite, were minimized. The subsamples were randomly chosen from a grid (0.3 by 0.6 m) that was laid on the level sampling surface. This grid was divided into eighteen one-square-decimeter intervals, numbered one through eighteen. Seven numbers were chosen between one and eighteen from a random-numbers table for each sample. The seven numbers so chosen were the collecting locations within the grid. Only three of these subsamples were used in this study. The remaining four are stored at the Department of Geological Sciences, SUNY at Binghamton.

One slab of rock (3 to 5 cm thick) was collected adjacent to each of the bedding-unit samples for preparation of thin sections perpendicular to stratification. These slabs were coated, in the field, with epoxy and tape to prevent disaggregation.

Laboratory Studies

Two or three thin sections of each bedding-unit sample (43 overall) provided the petrographic data. The percent composition of each thin section was determined by visual estimates using standard charts. The identities of the minerals observed in thin section were confirmed by X-ray analysis. Thin sections perpendicular to stratification were not used for the composition estimates because they had too much lateral variability in thickness. This variability was caused by the epoxy, which proved to be harder than most of the rock constituents, that filled the numerous partings during thin-section preparation. However, the thin sections perpendicular to stratification did supply supplemental information on the micropeloids.

A calcium-carbonate concretion and some adjacent clayshale were collected (this is a second set of samples; these were not studied petrographically) in order to study the ferrous iron content of chlorites. The clayshales were first ground with a mortar and pestle, and then sieved to separate and collect the size fraction finer than 21 um. The chlorites within the concretion were concentrated by removing the calcite with several treatments of dilute acetic acid. The remaining insoluble residues were then washed five times with distilled water. All of these fractions were suspended in distilled water and disaggregated ultrasonically for 10 minutes. Pipette analysis was used to collect the size fraction finer than 2 um (Folk, 1968, p.37). Centrifuging removed this fraction from suspension, and the resulting residue was resuspended in 10 ml of distilled water to form a concentrated solution. Slides of oriented clay minerals were made by placing 2 ml of the solution on glass slides and allowing them to air dry. These samples were then analyzed with X-ray diffraction at a scanning speed of 20° 20/min. from 4° to 70° 20. Peak intensities were measured for the first four basal reflections of chlorite. Two points on opposing peak shoulders, 0.2° 20 to each side of the peak center, were connected with a straight line. The length of the peak center line between the base level and the intersection with the line mentioned above was measured and termed peak intensity (Appendix 2). The proportions of Mg^{+2},Fe^{+2}, and Al^{+3} within the octahedral layers of chlorites were calculated by converting (using Lorentz factors) peak intensities into structure factors, and then following the method outlined by Brindley and Gillery (1956). Changes in peak intensity at 6.2° 20, 12.5° 20, 18.8° 20, and 25.2° 20 are all attributable to chlorite because no kaolinite was present in these samples.

Based on data from diffractograms (Appendix 2), the relative proportions of Mg^{+2}, Fe^{+2}, and Al^{+3} within the octahedral sites of chlorite were calculated for all of these samples (Table 2). The F(002)/F(001) ratio was used in calculating the proportion of Fe^{+2} because these two basal reflections are large amplitude peaks. Other structure-factor ratios involving the third-order and fourth-order basal reflections could also have been used, but these peaks are of small amplitude and are probably less reliable.

APPENDIX 2
X-RAY DIFFRACTION DATA

Sample	Basal Spacing	d-spacing	20	Peak Height (mm)	Lorentz Factor	Structure Factor SF=PH/LF
0.0cm	(001)	14.14A	6.25	20.9	9.024	1.52
	(002)		12.50	26.1	4.310	2.46
	(003)		18.80	8.9	2.668	1.83
	(004)		25.23	14.8	1.822	2.85
0.0cm	(001)	14.44A	6.12	15.0	9.221	1.28
	(002)		12.59	23.3	4.275	2.33
	(003)		18.74	2.4	2.678	0.95
	(004)		25.11	2.7	1.834	1.21
0.0cm	(001)	14.14A	6.25	7.4	9.024	0.91
	(002)		12.50	15.2	4.310	1.88
	(003)		18.80	2.8	2.668	1.02
	(004)		25.23	1.8	1.822	0.99
0.5cm	(001)	14.16A	6.24	16.1	9.039	1.33
	(002)		12.50	29.8	4.310	2.63
	(003)		18.81	3.2	2.666	1.10
	(004)		25.08	8.8	1.837	2.19
1.0cm	(001)	14.14A	6.25	25.3	9.024	1.67
	(002)		12.42	42.9	4.341	3.14
	(003)		18.79	3.1	2.670	1.08
	(004)		25.24	19.8	1.821	3.30
1.0cm	(001)	14.12A	6.26	17.1	9.019	1.38
	(002)		12.51	36.7	4.306	2.92
	(003)		18.89	8.9	2.652	1.83
	(004)		25.23	14.3	1.822	2.80
2.3cm	(001)	14.26A	6.20	13.9	9.099	1.24
	(002)		12.51	33.8	4.306	2.80
	(003)		18.74	7.1	2.678	1.63
	(004)		25.16	16.8	1.829	3.03

Sample	Basal Spacing	d-spacing	20	Peak Height (mm)	Lorentz Factor	Structure Factor SF = PH/LF
3.0cm	(001)	14.42A	6.13	12.8	9.206	1.18
	(002)		12.54	30.7	4.295	2.67
	(003)		18.85	5.9	2.659	1.49
	(004)		25.24	6.4	1.821	1.87
3.5cm	(001)	14.01A	6.31	24.7	8.936	1.66
	(002)		12.50	49.0	4.310	3.37
	(003)		18.63	16.9	2.698	2.50
	(004)		25.10	15.7	1.835	2.93
3.5cm	(001)	14.19A	6.23	13.0	9.054	1.20
	(002)		12.63	32.0	4.260	2.74
	(003)		18.86	7.9	2.657	1.72
	(004)		25.10	11.8	1.835	2.54
5.0cm	(001)	14.23A	6.21	13.3	9.084	1.21
	(002)		12.55	19.2	4.291	2.12
	(003)		18.76	3.8	2.675	1.19
	(004)		25.21	10.0	1.824	2.34
6.0cm	(001)	14.14A	6.25	6.9	9.024	0.87
	(002)		12.50	14.8	4.310	1.85
	(003)		18.74	0.8	2.678	0.55
	(004)	25.10	5.0	1.835	1.65	
6.0cm	(001)	14.12A	6.26	9.8	9.019	1.04
	(002)		12.51	16.7	4.306	1.97
	(003)		18.84	1.0	2.661	0.61
	(004)		25.10	6.1	1.835	1.82
7.5cm	(001)	14.23A	6.21	10.0	9.084	1.05
	(002)		12.53	19.8	4.294	2.15
	(003)		18.86	2.0	2.657	1.06
	(004)		25.24	10.2	1.821	2.37
22.6cm	(001)	14.37A	6.15	26.6	9.175	1.70
	(002)		12.45	51.3	4.329	3.44
	(003)		18.75	8.0	2.677	1.73
	(004)		25.13	11.3	1.832	2.48
24.0cm	(001)		6.26	13.0	9.019	1.20
	(002)		12.53	44.4	4.294	3.22
	(003)		18.75	4.2	2.677	1.25
	(004)		25.20	10.7	1.832	2.42

STRUCTURE-FACTOR RATIOS

Sample	F(002)/F(001)	F(002)/F(003)	F(004)/F(002)
0.0cm	1.62	1.35	1.16
0.0cm	1.83	7.80	0.52
0.0cm	2.07	1.83	0.53
0.5cm	1.97	2.40	0.83
1.0cm	1.88	2.92	1.05
1.0cm	2.12	1.59	0.96
2.3cm	2.27	1.72	1.08
3.0cm	2.27	1.79	0.70
3.5cm	2.03	1.35	0.87
3.5cm	2.29	1.59	0.93
5.0cm	1.75	1.77	1.11
6.0cm	2.19	3.39	0.89
6.0cm	1.89	3.21	0.93
7.5cm	2.05	2.02	1.10
22.6cm	2.02	1.99	0.72
24.0cm	2.68	2.57	0.75

Petrology, Taphonomy and Sedimentary Environments of Pyritic Fossil Beds from the Hamilton Group (Middle Devonian) of Western New York

Vincent B. Dick and Carlton E. Brett
Department of Geological Sciences
The University of Rochester

ABSTRACT

Several widely traceable beds in the Middle Devonian (Givetian) Hamilton Group (Ludlowville and Moscow Formations) of western New York State contain well preserved pyritic fossils and nodules. Consistent lithologic, paleontologic and preservational characteristics of all horizons suggest deposition under a particular, recurring set of environmental conditions.

The pyritic horizons occur in moderately fossiliferous, gray mudstones, transitional between black, fissile shales with low diversity pelagic faunas ("Cleveland" facies), and lighter gray, calcareous shales with high diversity benthic fossil assemblages ("Moscow" facies). Thus, Hamilton Group pyritic fossil horizons appear to represent a facies of the "dysaerobic zone" of a density stratified basin and not anoxic shale facies, as frequently assumed.

Preservation of uncompressed pyritic molds of shells and sponges, and wood with cell structures completely preserved demonstrate that void spaces were infilled by pyrite prior to substantial sediment compaction. Abundant pyritic steinkerns of molluscan shells and the occasional replacement of mollusk shell conchiolin by pyrite further indicate that substantial pyritization occurred prior to aragonite dissolution. Most pyrite occurs in semi-enclosed cavities of shells, skeletal pores or burrows. Pyritic fossils are relatively free of mud filling and contain, at most, four phases of pyrite: 1) shiny, finely crystalline and commonly stalactitic liner pyrite occurs as rims on or in shells and skeletal pores; 2) microcrystalline sedimented pyrite; 3) a related calcitic pyrite both probably originally loose framboidal material, form geopetal fillings in shell cavities; and 4) macrocrystalline, often bladed, overpyrite encrusts the outer surfaces of some skeletons and burrows and forms small spheroidal and irregular nodules. Liner pyrite formed the earliest phase and commonly was shattered during compaction. Likewise sediment pyrite remained unconsolidated during early diagenesis; it was later indurated, either by calcite (calcitic pyrite) or later pyrite cements. Overpyrite formed later than other phases, during and following compaction. Other shell-filling minerals, including calcite spar, barite and galena, represent later diagenesis.

Pyrite formation probably began in the upper few centimeters of unconsolidated sediment, within reducing microenvironments afforded by skeletal cavities and lined burrows. Decaying organic matter at these sites localized sulfate reduction and established concentration gradients which aided in mobilization of ferrous iron and sulfate to isolated nuclei.

Biostratinomic evidence indicates that sediments containing pyrite aggregates were deposited in low energy, oxygenated waters, and were well bioturbated, even though the redox potential discontinuity (RPD) probably lay not greater than 5 cm below the sediment-water interface. Discrete beds of pyritic nodules and enrolled trilobites, complete sponges and articulated echinoderms indicate episodic rapid burial and associated elevation of the RPD within the mud. Such burial events emplaced organic matter into generally organic-poor sediment and may have initiated iron sulfide precipitation.

INTRODUCTION

Beds of pyritic fossils and nodules, typically associated with dark-gray marine mudstone facies, have been documented at various levels of the geologic column, including the Devonian Silica Formation of Ohio (Nussman, 1975), the Mississippian Fayetteville Shale of Arkansas (Zangerl and others, 1969), and the Jurassic Posidonienschiefer of Germany (Einsele and Mosebach, 1955), and Oxford Clay of Great Britain (Brown, 1966; Hudson and Palframan, 1968; Hudson, 1978, 1982; Morris, 1980).

Pyritic fossils and nodules occur in several thin, widespread horizons in the Middle Devonian Hamilton Group (Fig. 1) in western New York State (Loomis, 1903; Fisher, 1951; Beerbower and others, 1969; Brett, 1974), particularly in the Ludlowville and Moscow formations of the upper Hamilton Group. Although previous attempts have been made to understand their mode of formation, few studies have considered the

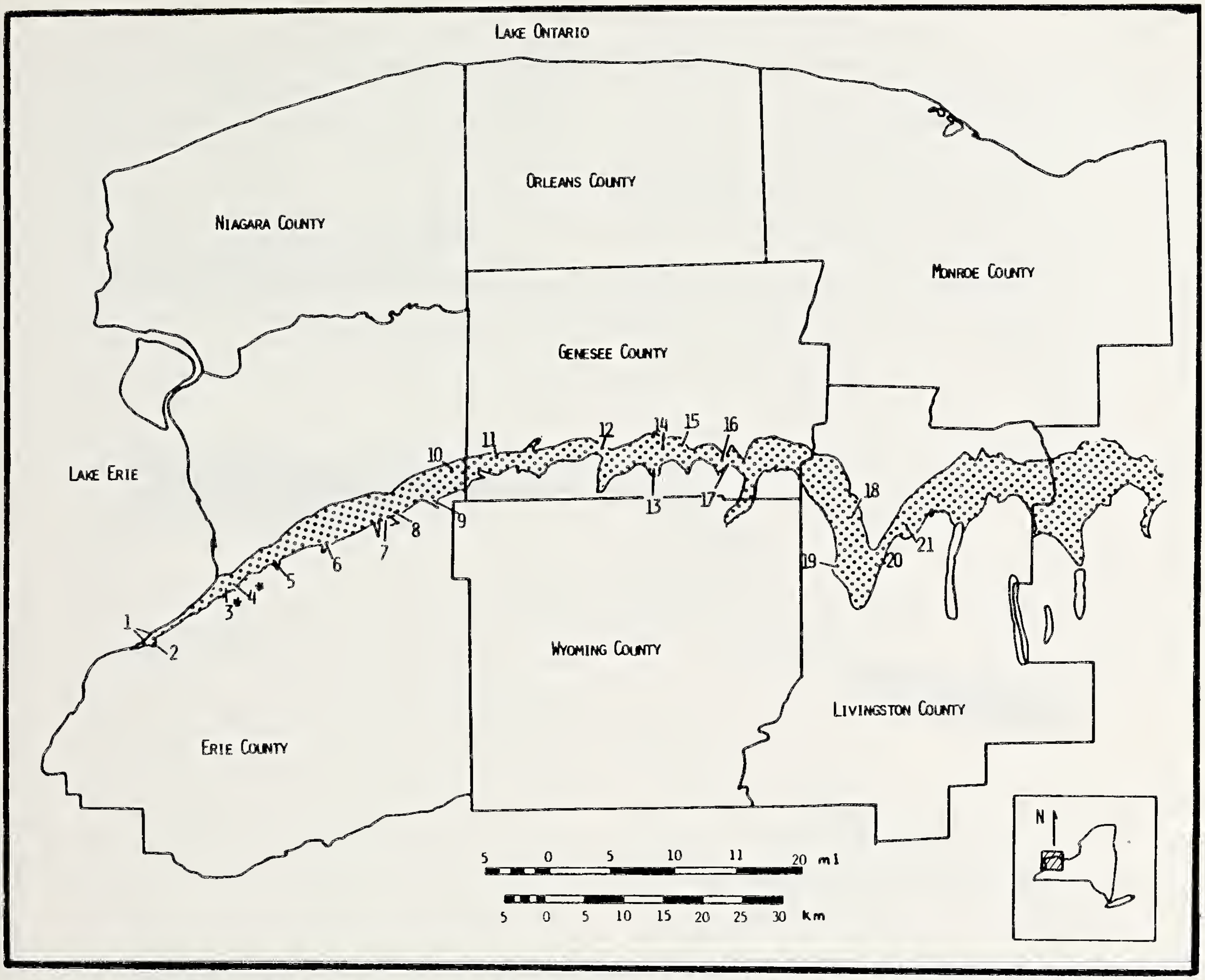


Figure 1: Outcrop belt of the upper Hamilton Group (Ludlowville and Moscow Formations) in western New York State. Numbered sites correspond to measured and sampled localities within the outcrop belt. Locality descriptions are listed in Appendix I.

stratigraphic and geographic distribution of pyrite horizons, or their relation to associated sediments. To date, only three beds have been studied in any detail; the Leicester Pyrite at the top of the Hamilton Group (Clarke, 1885; Loomis, 1903; Kalliokowski, 1966; Izard and Clemency, 1967; Park and Weiss, 1972), the Ledyard Shale Member "Alden pyrite" (Fisher, 1951; Izard and Clemency, 1967; Domagala and Selznick, 1979), and isolated pyritic patches from the Wanakah Shale (Beerbower and others, 1969).

At least six additional pyritic fossil beds occur in the Hamilton Group (Figs.1 and 2); one in the lower Wanakah Shale Member (Ludlowville Formation) and five in the Windom Shale Member (Moscow Formation) (see Dick, 1982 for detailed discussion).

Pyritic beds appear to be closely facies-controlled. The fossil assemblages associated with these beds are similar in composition and distinct from those of other Hamilton beds at all levels. Published faunal lists and preliminary fieldwork revealed that assemblages in all eight beds consist of a few species of small brachiopods, nuculoid bivalves, various archeogastropods, orthoconic nautiloids, goniatites, and the trilobites *Greenops* and *Phacops* (Brett and others, in press). Many of the benthic forms are diminutive (Loomis, 1903; Fisher, 1951), suggesting stressful environmental conditions (Cloud, 1948; Hallam, 1965), although examples of true stunting are probably rare. The faunal assemblages also are distinct from previously described Hamilton Group assemblages such as the *Tropidoleptus, Ambocoelia, Leiorhynchus* and rugose

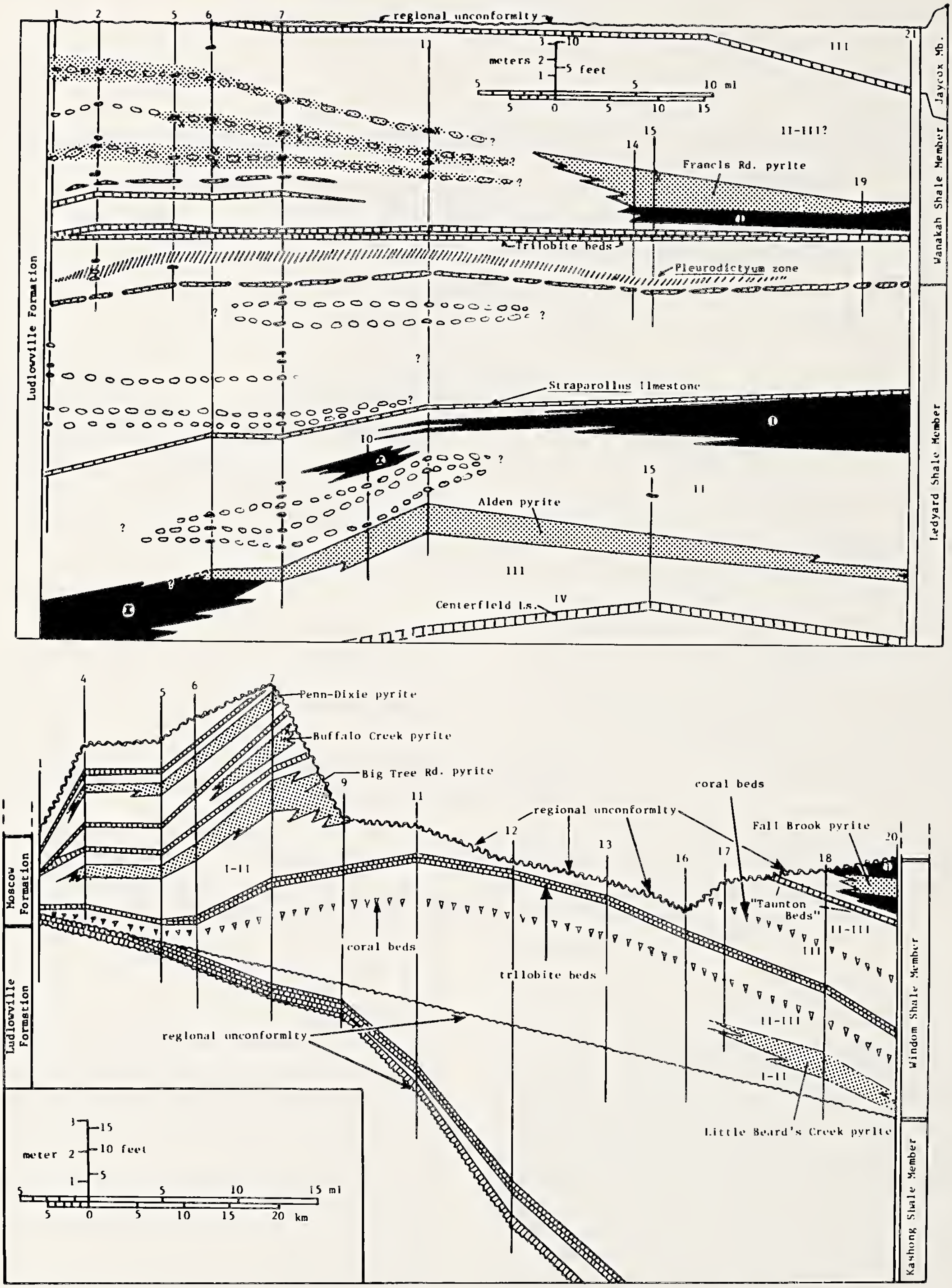


Figure 2: Stratigraphic relationships of upper Hamilton Group pyrite beds. Roman numerals refer to litho- and biofacies types I – IV described in the text; black shading = black shale lithofacies; stipple pattern = gray, pyritic mudstones. X's identify stratigraphic horizons of pyritized fossils and nodules; ellipses = carbonate concretions.
A. Ludlowville Formation; Ledyard and Wanakah Shale Members and pyrite beds.
B. Moscow Formation; Kashong and Windom members.

coral faunas, (see Cleland, 1903; Williams, 1903; Cooper, 1930; Grasso, 1970; McCollum, 1980). Indeed, it is suggested by Fisher (1951), Brett (1974), and Brett and others (in press) that the formation of pyritic beds and the unique faunas associated with them are related in occurrence.

Pyritic beds are distributed non-randomly in stratigraphic sections of the upper Hamilton Group (Fig. 2). In all examples studied, the pyritic assemblages occur in the transitional beds between dark-gray and black, fissile shaley facies, with primarily pelagic fossils ("Cleveland" facies of Caster, 1934; "Marcellus" facies of Rickard, 1981), and lighter gray calcareous shales with abundant benthic fossils ("Moscow" facies of Grasso, 1970). This stratigraphic position agrees with the developing concept that the northern Appalachian Basin was a density stratified, euxinic basin during Middle to Late Devonian time (Sutton and others, 1970; Byers, 1977; Cluff, 1980; McCollum, 1980). Pyritic beds of the Hamilton appear to characterize gray mudstone facies of the upper dysaerobic zone of Rhoads and Morse (1971) (or lower aerobic zone, O_2 levels $>$ 0.5 ml/l; Thompson and others, 1985) but not, as frequently assumed, the anoxic black shale facies.

Fossils from pyritic beds are excellently preserved as partial to complete steinkerns of pyrite. This includes originally aragonitic molluscan shells which are poorly preserved in most other Hamilton beds. Pyrite mold formation thus appears to occur prior to, or, in part, contemporaneously with the earliest stages of compaction and aragonite dissolution. Geochemical studies of the last decade have demonstrated that sedimentary pyrite formation, although not fully understood, is a fairly rapid, ubiquitous process in fine-grained marine sediments (Berner, 1964-1981; Kaplan and others, 1963; Curtis and Spears, 1968; Rickard, 1975; Howarth, 1979). Indeed, recent discovery of shell replacement by pyrite in living marine bivalves (Clarke and Lutz, 1980) indicates that some pyritic preservation occurs prior to strictly defined temporal limits of taphonomic processes.

The purpose of the present paper is to provide a detailed description of pyritic fossils and nodules from the Hamilton Group and to develop a depositional and diagenetic model for pyrite beds based on their position in the facies spectrum of the Hamilton.

STUDY AREA AND METHODS

Exposures of pyritic units in the Ledyard and Wanakah Shale Members (Ludlowville Formation) and the Windom Shale Member (Moscow Formation) of the upper Hamilton Group were examined at twenty-one localities. General stratigraphic relations of these units and the remainder of the Hamilton Group in the study area are illustrated in Figure 2; pyritic horizons are described in detail in Dick (1982) and Brett and Baird (1982).

Detailed sections were measured at all localities to determine lateral and vertical stratigraphic variability of the pyritic beds. Precise measurements also were made within pyrite bearing intervals to determine internal variation.

Fossils and nodules were sampled at four localities, two each for the Windom and Ledyard Shale members. Exposures of approximately 1 m^2 areas were cleared at each of these sites and fossil/nodule counts made through 10 cm intervals (i.e., from a volume of $100 \times 100 \times 10$ cm $= 10^5$ cm^3). Sampling at 10 cm levels was repeated for each horizon through 1.5-2.5 m vertical sections. Orientations of concavo-convex shells, azimuthal (compass) orientation of elongate objects (e.g., cephalopods), shell articulation/disarticulation ratios and the nature of the enclosing rock matrix (color, fissility, degree of lamination, burrow structure) were all recorded at sampling localities. For the purpose of mass balance calculations, an accurate record of bulk shale sample and total pyrite mass was kept. At the remaining localities faunal presence-absence surveys were conducted and variations in the nature of the pyritic beds were noted.

Polished sections and thin sections of oriented pyritic fossils and nodules were used for petrographic study. Sections were etched with 5% HCl and stained, using potassium ferricyanide alizarin red-S stain (after Dickson, 1966) to enhance carbonate fabric. X-ray diffraction was used to verify identification of selected mineral phases.

GEOLOGIC SETTING

The Hamilton Group is a sequence of clastic and thin carbonate sediments deposited primarily in shallow offshore and nearshore marine settings (Cooper, 1930, 1957; Rickard, 1981). The present study is concerned with the upper portion of the Hamilton Group (Ludlowville and Moscow formations) in western New York.

Within the study area (Fig. 1) the Hamilton Group represents sediments deposited in a northern arm of the Devonian Appalachian Basin (see Savarese et al., this volume). This inland sea was flanked on the east and southeast by the developing Catskill Delta complex and on the west and north by very low relief cratonic shelf regions (Dennison and Head, 1975). Deepest portions of the basin lay to the south and southeast of the study area (Dennison and Head, 1975). The Hamilton Group has, in the past, been interpreted as a rather simple basin-filling sequence, resulting from a fluctuating supply of clastic fines from the prograding Catskill Delta complex to the southeast. Recent studies demonstrate that many thin faunally and sedimentologically distinctive units, including the pyritic shell fossil beds discussed herein, are widely traceable along the east-west trending Hamilton outcrop belt. Indeed, certain of these units even crosscut north-south oriented facies boundaries, established on the basis of lithology alone. These observations indicate a more important control on the distribution of fossils and certain other sedimentary features than simply sediment supply or grain size. Such controls probably include oxygen levels, turbidity, turbulence, and other depth-related

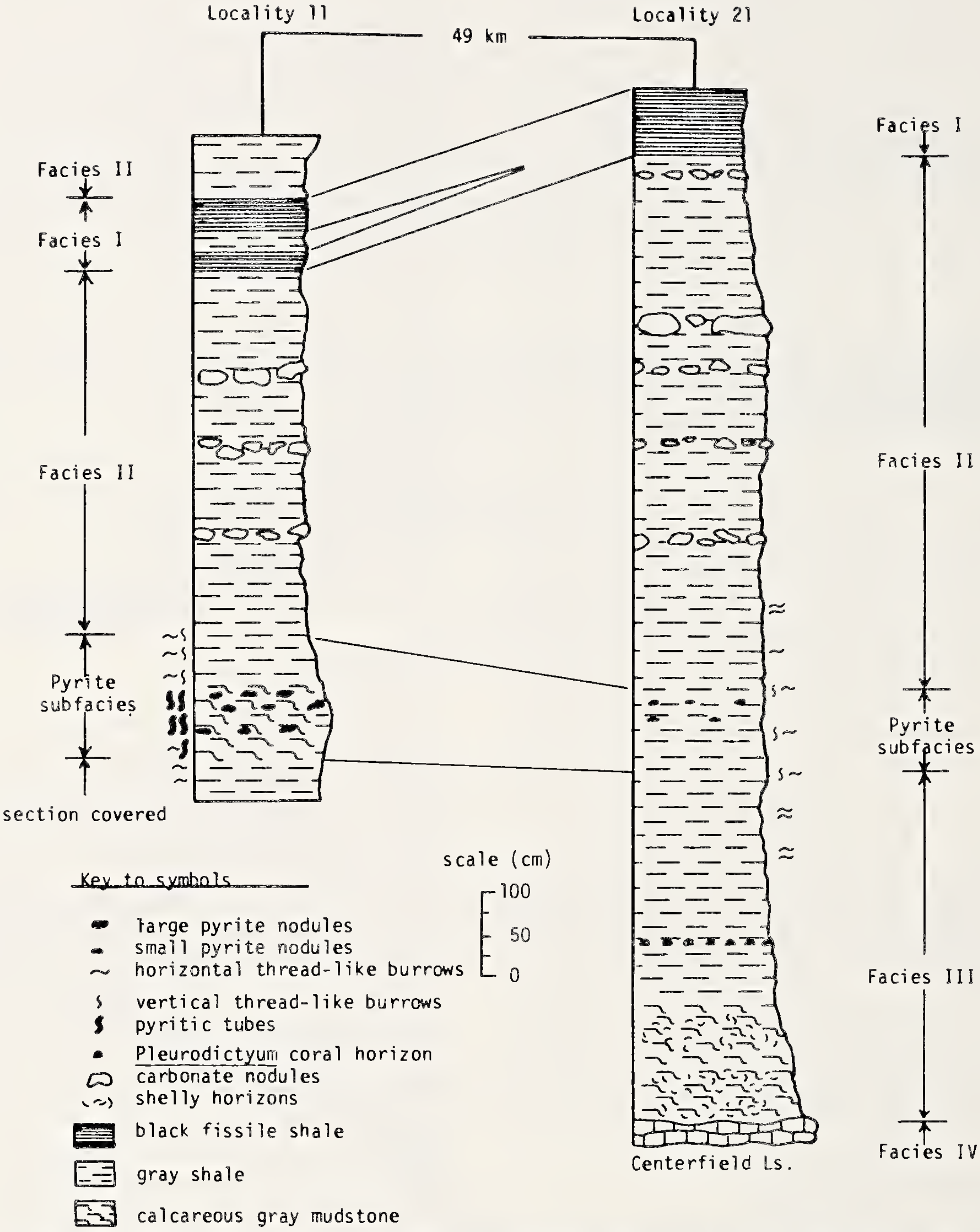


Figure 3: Stratigraphic columns of the Alden pyrite from Elevenmile Creek (loc. 11) and Jaycox Creek/Wheeler Gully (loc. 21). Each column shows the vertical transitions of facies types from Facies I or IV through the pyritic interval. Comparison of the two localities shows the lateral changes of trace fossils and sediment fabric.

parameters, all of which may be evaluated directly, or by inference from careful and rigorous examination of the rock units.

Four general depositional environments may be identified in the upper Hamilton Group. They are represented by four litho- and biofacies types (Fig. 3) that range from dark, fissile shales deposited under conditions least conducive to benthic marine life (facies I), to calcareous shales, biostromes and limestones deposited under highly favorable conditions, allowing development of a rich marine fauna (facies IV; see Dick, 1982; and Brett et al., 1983, for full discussion of all facies types). Facies I

UNIT			NUMBER CONCAVE-UP	TOTAL NUMBER	PERCENT UNSTABLE	
LEDYARD SHALE	Alden pyrite	Locality 11	171	419	41%	D
			152	243	63%	A
			14	52	27%	L
		Locality 15	166	298	56%	D
			326	648	50%	A
			37	60	62%	L
WINDOM SHALE	Penn-Dixie and Big Tree Road pyrites	Locality 3	147	302	49%	C
		Locality 4	30	56	54%	D
			411	774	53%	A
			4	6	66%	T
			167	356	47%	C
	Little Beard's pyrite	Locality 18	174	274	64%	C

Table 1: Concave up/down orientation data for four pyritic intervals from the Ledyard and Windom Shales. Data collected from:

D = *Devonochonetes*
A = *Ambocoelia*
L = *Longispina*
T = *Tropidoleptus*
C = **Combined**

and II contain the primary concentrations of pyrite in the Hamilton Group and were used for the comparative study described herein.

Facies I consists of brittle, black to olive-gray shale composed primarily of illite and chlorite (Towe and Grim, 1963). These beds contain abundant organic walled cysts (Wygant, this volume) and authigenic pyrite in the form of disseminated framboids that often give the shale a rusty stain upon weathering. Carbonate content is low and concretion occurrence is rare. Well developed fissility, coupled with a rarity of trace fossils and low diversity of benthic body fossils, reflects a relatively quiet, poorly oxygenated bottom setting at the time of deposition. Body fossil remains present are most frequently those of planktonic or nektonic forms (e.g., *Styliolina, Tornoceras*). Preservation is always poor, with shells frequently highly compressed, or displaying signs of shell carbonate dissolution, and little or no pyritic steinkern preservation.

Facies II mudrocks are higher in carbonate content than are those of facies I and range from slightly fissile shales to blocky, bioturbated mudstones. The mudstones are generally soft and medium gray but have a clay mineralogy quite similar to facies I (Towe and Grim, 1963). Bulk organic content is lower than in facies I but fossils and concretions are both considerably more abundant. The presence of moderately well developed benthic infaunal and epifaunal communities and the bioturbated fabrics reflect better oxygenation at the time of deposition, both within the water column and in the upper sediment. Occasional shell stringers and gutter casts indicate episodic bottom currents over an otherwise low energy depositional setting.

The moderately diverse fauna of facies II includes the nektonic and planktonic forms of facies I as well as abundant semi-infaunal nuculoid bivalves, small epifaunal brachiopods (chonetids, ambocoeliids and small *Tropidoleptus*), and various archeogastropods (e.g., *Palaeozygopleura, Glyptotomaria*). The trilobites *Greenops* and *Phacops* are abundant and frequently occur in horizons of dominantly enrolled individuals. Trace fossils, represented by vertical to slightly inclined pyrite-filled tubes, are abundant. Preservation of body fossils is predominantly by pyrite infilling and/or encrustation by pyrite crystals.

BIOSTRATINOMY AND SEDIMENTOLOGY OF PYRITIC FOSSIL BEDS

The biostratinomy of fossils found associated with pyritic beds provides sedimentologic data for interpretation of depositional regimes. The following sections describe sedimentologic features of the matrix rock and the use of trace and body fossils as a means of assessing bioturbation, bottom oxygenation, and sedimentation patterns.

Shell Breakage, Disarticulation and Orientation

The degree of post-mortem reworking of skeletons may be qualitatively assessed by observing the amount of abrasion and breakage present in fossil shells. Shell material in all pyritic units studied is excellently preserved. Calcitic shells, such as brachiopods, occasional bryozoans (e.g., Alden pyrite at Loc. 11) and pelmatozoan columnals, are generally unabraded. Most bivalved shells are articulated and closed; this includes shells of pelecypods which elsewhere in Hamilton Group units are frequently found splayed open or disarticulated. Certain delicate shell features such as chonetid hinge spines are commonly preserved intact.

Such observations suggest that little particle movement and contact affected these shells, implying that currents operating at the time of deposition were weak to non-existent, or that shell layers were subjected to rapid burial. Concavo-convex shells, and azimuthal orientations of elongate objects provide quantitative data for the existence of paleocurrents (Brenchley and Newall, 1970; Reyment, 1971), and a check on qualitative observations of post-mortem reworking.

Results of concave up/down data for brachiopods of the Ledyard and Windom shales are summarized in Table 1. These brachiopods generally are small (0.5-1.0 cm), lightweight, concavo-convex or plano-convex shells. In adult life they probably lived unattached with the convex surface of the shell resting on the sediment, maintaining the commissure away from the substrate (Rudwick, 1970; Bowen et al., 1974). Due to their hydrodynamic properties, these shells were prone to post-mortem inversion to concave-downward orientations in the presence of even weak currents, especially on mud substrates (Brenchley and Newall, 1970). Thus, they are good indicators of current activity.

Statistically non-random orientations should result from total lack of post-mortem disturbance (all shells in concave-up "life" position), or from prolonged or strong current disturbance (all shells flipped to preferred concave-down stable position — see Reyment, 1971). The average "percent unstable" (i.e.,concave-upward) for each horizon in Table 1 is about 53%, thus falling between the non-random extremes described above. This implies some disturbance which may have resulted from a gentle periodic current, bioturbation and/or predation. Bioturbation is a very probable cause and is discussed below.

Rose diagrams were constructed from azimuthal orientations of elongated objects including orthoconic nautiloids, complete trilobites and wood fragments. Orientations, summarized in Figure 4, provide directional information on currents operating. Modality of each horizon was tested by calculating mean vector direction (e), vector magnitude (r) and mean angular deviation (s). These parameters appear with each rose diagram (Fig. 4). Truly polymodal samples should have a high mean angular deviation (Reyment, 1971). All pyrite beds sampled exhibit polymodal distributions and no apparent preferred alignment exists. Since data were taken from a number of levels within each interval to produce one rose diagram, each diagram is from a time-averaged sample. Therefore, the data do not rule out gentle periodic current action, but argue against anything greater.

Bioturbation

Bioturbation destroys original sediment lamination created by the settling of clay particles and thereby also inhibits the development of a fissile fabric (Byers, 1974; Potter et al., 1980). Thus, when bioturbation increases, fissility of shales decreases.

Dark-gray shales adjacent to all pyritic beds have moderate to well-developed fissility. This implies quiet, slow sedimentation and, more importantly, lack of infaunal organisms. Pyrite beds, in contrast, typically have abundant pyritic tubular structures and relatively non-fissile, bioturbated fabric, implying the existence of an active infauna.

Although no formal ichnogeneric assignment has been given to the types of pyritic trace fossils found within Hamilton strata, ichnofossils that are at least similar in form to those described here have been documented in the literature (see Rhoads, 1970; Frey, 1970, for review).

The form herein referred to as "thread-like burrows" is most similar to the ichnogenus *Trichichnus* of Frey (1970). Hamilton thread-like burrows are long (tens of centimeters), unbranched and sinuous. They are cylindrical (diameter about 1.0 mm) and predominantly horizontally oriented (Fig. 5A).

Pyrite structures of the Hamilton Group referred to as "tubes" are quite similar to larger "mineral-filled burrows" of Frey (1970). They are long, cylindrical and robust (1-2 cm width × 5-30 cm length), usually unbranched, and are variously oriented with single specimens sometimes being both vertical and horizontal along their length (Fig. 5A). Frey (1970) indicates that these structures probably do not represent a single ethological category, although their form may be closely allied

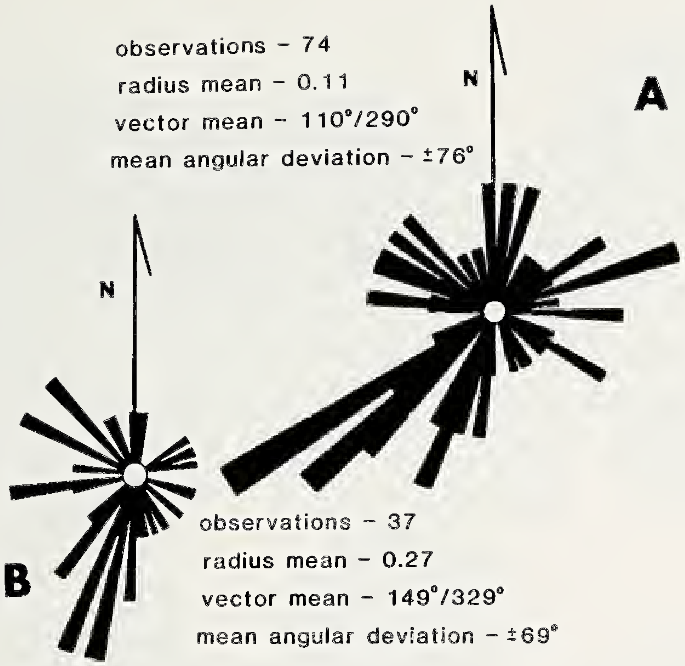


Figure 4: A. Azimuthal orientations of nautiloid cephalopods and trilobites from the Ledyard Shale, Alden pyrite.
B. Azimuthal orientations of nautiloid cephalopods and wood axes from the Windom Shale.

to burrows constructed by the Recent polychaete *Clymenella*. If this is true, then these are dwelling tubes (domichnia) and vertical tube length may reflect the distance from the sediment-water interface to feeding layers within the sediment, usually below the RPD (Aller, 1980). Pyritic tubes of the Hamilton Group average 3.5 cm in length (± 1.3 cm), suggesting that the anoxic boundary lay at about this depth within pyritic interval sediments. Furthermore, these dwelling tubes are the type built by sedentary suspension feeders in very unstable, slightly oxygenated, water-rich substrates (Rhoads, 1970; Aller, 1978).

Recently, Thompsen and Vorren (1984) described pyritic tubes of comparable size and morphology from late Pleistocene and Holocene marine sediments near Norway. In this example, early diagenesis of pyrite occurred in reducing microenvironments around mucous-lined polychaete burrows in an otherwise slightly oxygenated sediment.

Various body fossil forms also strongly suggest the existence of soft, thixotropic substrates within the pyritic intervals of the Hamilton Group. Nuculoid bivalves, for example, which are quite abundant in pyritic intervals (Table 1), have changed very little in form and (by inference) life habit since the Devonian (McAlester, 1968). Modern nuculoids inhabit muddy substrates of the Atlantic coast, burrow to 2-3 cm depth and vigorously rework sediment, sometimes to the point of exclusion of other benthic fauna (Rhoads, 1974). This type of life habit would be consistent with other observations made in this study that suggest the presence of a thixotropic substrate in pyritic intervals.

Other Hamilton Group fossils show adaptations that might have helped them compensate for substrate instability, such as decreased shell mass (e.g., thin shelled chonetids). Spinosity is also common (e.g., *Greenops*; see Babcock, 1982), a feature that allows organisms to "float" atop a shifting substratum (Thayer, 1975).

An actively mixed substrate, such as that suggested thus far, would not be expected to maintain shells in "life" orientations after death of the organism. This provides a mechanism to explain concave-down orientations of shells without invoking current action; the concave up/down ratios of Table 1 thus probably reflect partial bioturbation of predominantly concave-up life assemblages.

Nodule-Fossil Layers

Several pyrite-bearing intervals contain widely exposed layers of large nodules and/or particular fossil types. These horizons display similar taphonomic features that presumably resulted from similar types of burial environments.

For example, the Alden Pyrite bed (Fisher, 1951; Dick, 1982) is most easily recognized in outcrop by a prominent layer of large nodules, usually about 30-40 cm from the base of the

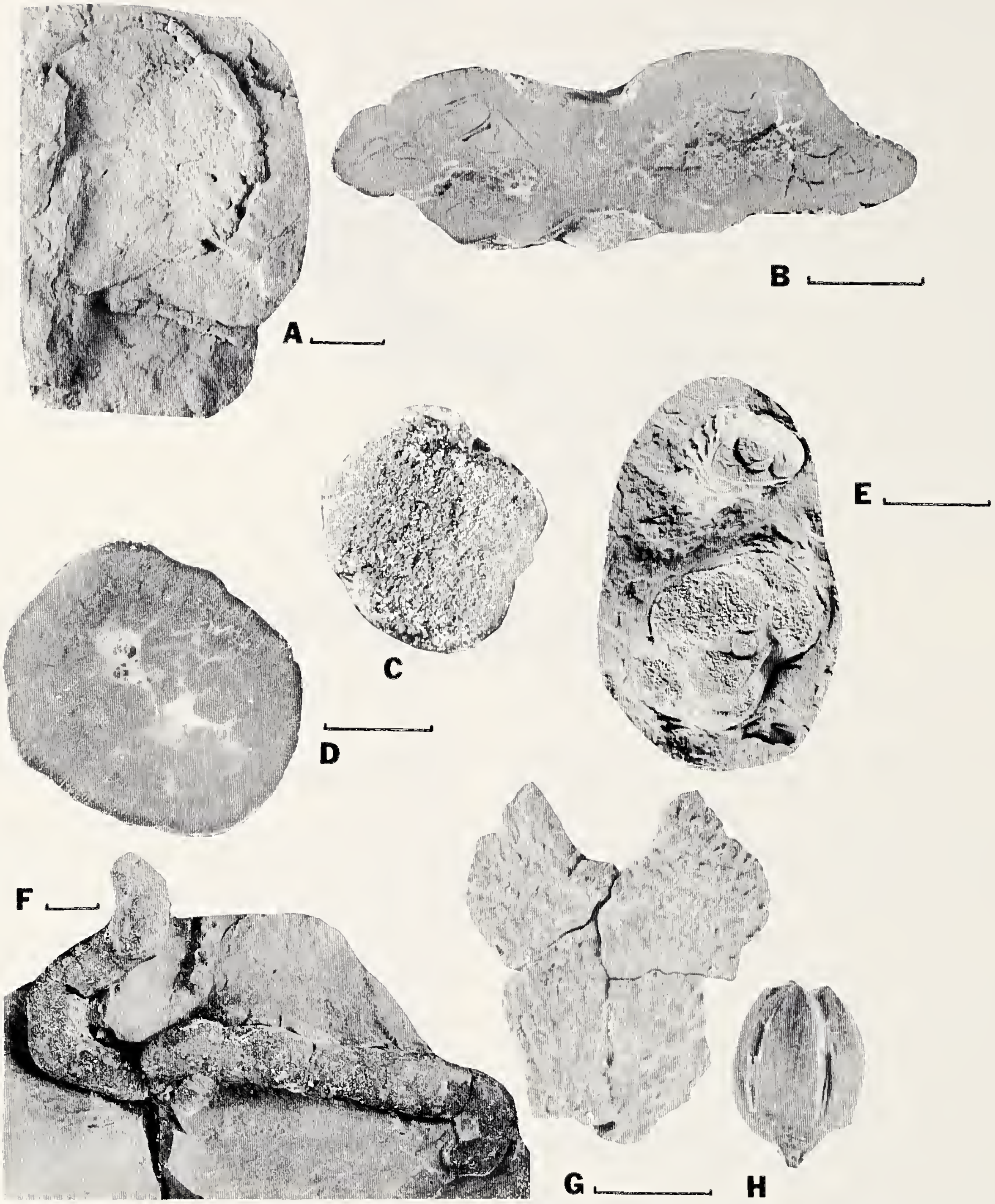


Figure 5: Forms of pyritic fossils/nodules contained in upper Hamilton Group pyrite beds. All scale bars = 1 cm.

A. Horizontal "thread-like" burrow; Ledyard Shale, shales immediately overlying the Alden pyrite bed; loc. 11.

B. Pyritized ellipsoidal nodule, probably a coprolite; note incorporated shelly debris, including shells of several small ammonoids; Ledyard Shale, Alden pyrite; loc. 15.

C., D. Exterior and polished interior views of two pyrite nodules, bearing no recognizable nuclei; note fractured interior filled with barite; Ledyard Shale, Alden pyrite; loc. 15.

E. Top view of sample surface (parallel to bedding) showing two enrolled specimens of the trilobite *Phacops rana*, partially encrusted with crystalline overpyrite; Ledyard Shale, Alden pyrite bed, loc. 11; RMSC 83. 70. 1.

F. Sinous pyritic "tube"; Ledyard Shale, Alden pyrite, loc. 15; RMSC 83. 71. 4.

G. Unidentified lithistid sponge; Windom Shale, Big Tree Road pyrite bed; loc. 4; RMSC 83. 66. 1.

H. Blastoid *Hyperoblastus*, permineralized by pyrite (now altered to limonite); Windom Shale, Big Tree Road pyrite bed; loc. 4; RMSC 83. 66. 15.

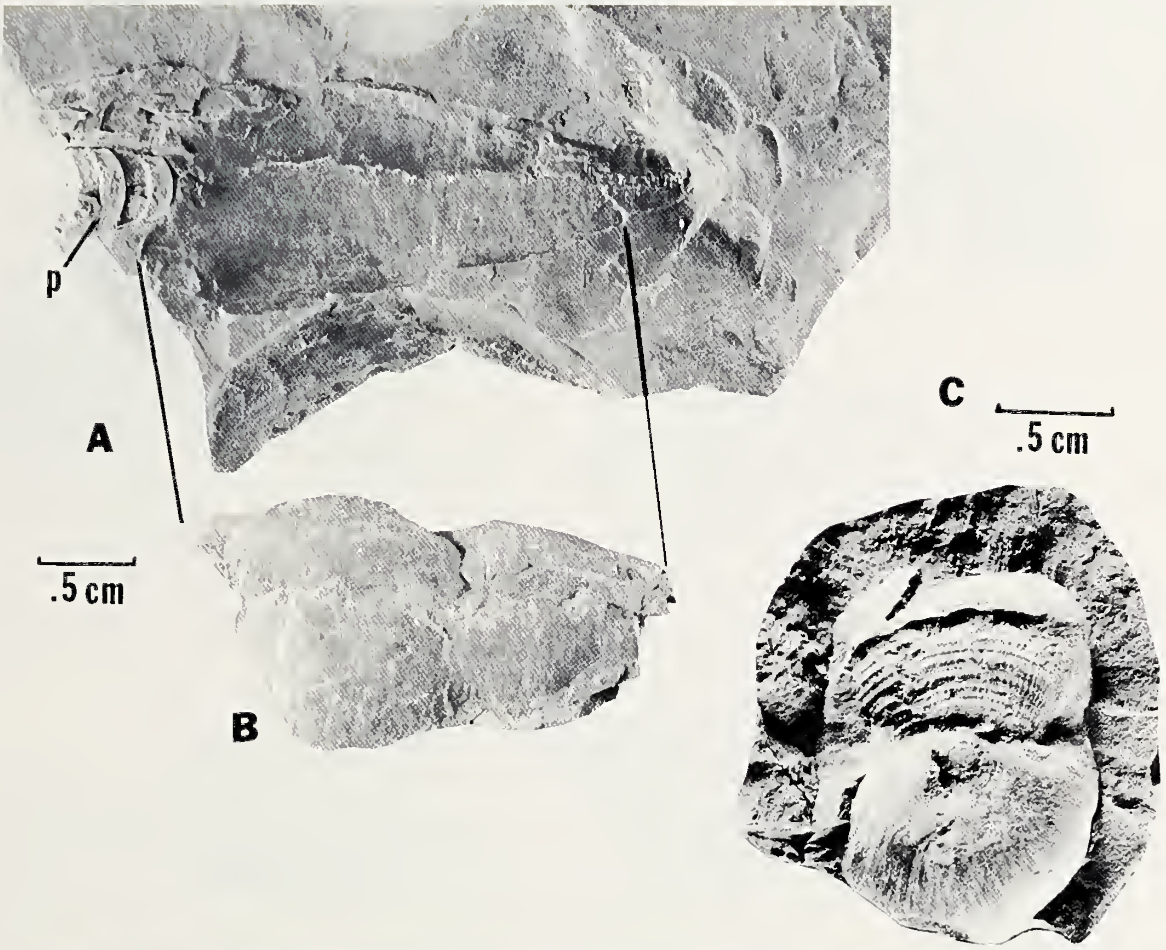


Figure 6: Variations in preservation of non-pyritized fossils from pyrite-bearing shale intervals.
A., B. Compressed mud steinkerns of nautiloid; note small amount of pyrite (p) that formed before mud filling and compaction around steinkern (B) giving bilobed, slickensided appearance; Ledyard Shale, Alden bed; loc. 11.
C. Compressed mud steinkern of the gastropod *Mourlonia itys*; note partial preservation of external ornamentation which has been impressed onto the internal mold; Ledyard Shale, Alden pyrite; loc. 15.

interval, seen at most localities (Figs.3, 5C-D). This layer may be traced across nearly 70% of the studied outcrop belt. Nodules commonly have nucleii composed of pyritic fossil steinkerns or burrow tubes. Others have a core of cubic or pyritohedral pyrite, enclosed by a layer of bladed overpyrite (see petrology section for mineral description). Such nodules may have formed around sites of decaying soft-bodied organisms or coprolites containing no hard shell parts. The similarity of these nodules in morphology, abundance and stratigraphic expression suggests that they result from a single event or phenomenon. Similar layers are present in three of the Windom pyrite beds, in addition to the Alden pyrite.

Fossil layers suggest more specific information about the character of these phenomena or events. For example, enrolled trilobites define horizons in the Ledyard and Windom pyrite beds. Fisher (1977; pers. comm.) has determined experimentally that certain semi-infaunal arthropods (*Limulus*, for example) will burrow into sediments and enroll, due to onset of stressed environmental conditions, especially decrease of dissolved oxygen. Localized layers of enrolled trilobites in an outcrop where trilobites are otherwise usually non-enrolled suggest temporally unique phenomena causing a behavioral response of the trilobites (Babcock and Speyer, 1983). Lateral persistence of fossil layers indicates that events may have been regional in effect, perhaps even defining localized time surfaces or isochrons (see paper by Brett and others, present volume). Causes of environmental stressing that fit these criteria might be decreased oxygen levels, salinity changes (associated with storm run off) or disturbance of anoxic bottom muds.

Bedding planes crowded with small specimens of one or a few species of brachiopods are often associated with thin, sheetlike pyritic crusts. Such assemblages have been described as isolated spat-fall communities preserved in situ due to mass mortality by rapid burial (Beerbower et al., 1969; Bray and Beerbower, 1969; Beerbower, 1981). Although the constant high sedimentation rate suggested by these authors is not supported by the present study, the idea of episodic rapid sedimentation causing death and/or burial of these communities is favored. The occurrence of well preserved sponges and articulated blastoids (Fig. 5G, H) strongly supports this interpretation. Firstly, the occurrence of stenotopic forms in itself indicates that, at times, background sedimentation rates and turbidity a few centimeters above the sediment water interface were at a minimum. Conversely, the excellent preservation of these delicate forms strongly suggests rapid burial immediately after (or causing) death, precluding disarticulation.

EARLY DIAGENESIS OF PYRITIC FOSSILS

Preservation of Pyritized vs. Non-Pyritized Fossil Assemblages

Comparison of fossils from Hamilton Group pyritic assemblages with those of other Hamilton non-pyritic assemblages provides strong evidence that the former underwent selective early diagenesis. In addition, comparison points up a number of examples of preservational bias between the two assemblages that result from differences in biostratinomic and early diagenetic processes. Notable distinctions between pyritic and non-pyritic fossil assemblages are discussed below in relation to steinkern constituents, compaction and aragonite dissolution, and trace fossil preservation.

Absence of Mud from Pyritic Steinkerns

Most non-pyritized fossil steinkerns in the Hamilton Group contain mud and minor late diagenetic calcite spar (Fig. 6). Mud probably was introduced into empty shells by draft-filling or by thixotropic injection. Sediment commonly infiltrated even the inner chambers of cephalopod shells and the inner whorls of gastropods.

Conversely, pyritic fossil steinkerns have little or no mud as an internal constituent, although, rarely, mud may partially fill the body whorls of gastropods or cephalopod living chambers. Hence, the formation of iron sulfides must have preceded mud infilling.

This suggests that pyritization began while the soft parts of the organism were still intact, blocking the aperture and preventing introduction of mud. Only the living chamber could be filled with mud, after decay of soft body parts. No instance of pyritization of soft parts has been found. However, one specimen of a nautiloid contains probable gut contents consisting of small goniatites in the living chamber.

Aragonite Dissolution and Compaction

The degree of fossil shell compaction provides a second distinction between pyritic and non-pyritic assemblages. Non-pyritized shells are mechanically distorted to varying degrees due to compaction. This is especially true of originally aragonitic shells which normally occur as compressed, internal molds (Fig. 6). Some three-dimensional form is preserved in these molds, apparently due to partial induration of infilling mud. This hardened mud, upon aragonite dissolution and subsequent compaction, offered resistance to clay particles sliding past, producing a slickensided surface on the compressed steinkerns. In contrast, calcareous shells with pyritic steinkerns display relatively little compactional deformation. Originally calcitic skeletons are well preserved and display infilling with pyrite; aragonitic shells occur as uncompressed, naked steinkerns, the aragonite having been dissolved. Hence, pyritization evidently predated both aragonite dissolution and most compaction. Most fracturing present on pyritic steinkerns occurred during or after pyrite formation, being possibly enhanced by internal pressure due to crystallization of pyrite (Figs.7A, 8E, F).

Microenvironments of Shells

Pyrite aggregates tend to occur only in skeletons with partly enclosed cavities, such as cephalopods (Figs.7, 8), gastropods, or enrolled trilobites; disarticulated elements such as single valves or trilobite sclerites commonly show little or no pyritization. This condition suggests that either 1) pyrite formed only in partially closed microenvironments, 2) pyrite diagenesis was favored by decaying soft-parts within the shells, or both. In any case, an early diagenetic origin of the pyrite is indicated.

Most notable in this regard is the selective pyritization of bivalves. Nearly all pyritized bivalves consist of the steinkerns of complete, tightly closed shells. Conversely, single valves or shells with valves widely splayed open, from the same beds, are generally not pyritized and occur as compressed molds as in non-pyritic beds. Bivalve shells tend to splay open and disarticulate rapidly upon death because of their hinge ligaments (Müller, 1979). Therefore, they would not seem likely to form tightly closed cavities; however, pyritic steinkerns are abundant in some species, e.g., nuculoids (Fig. 8E, F). The occurrence of pyritic steinkerns seems to be related to life habit. Twelve of the seventeen most common aragonitic bivalve species in the pyritic beds, are infaunal (Stanley, 1970). As such, members of these species often died within their burrows in sediment and pressure of those sediments prevented splaying of valves, allowing microenvironment formation. Conversely, the other five species, which are epifaunal, are rarely pyritized.

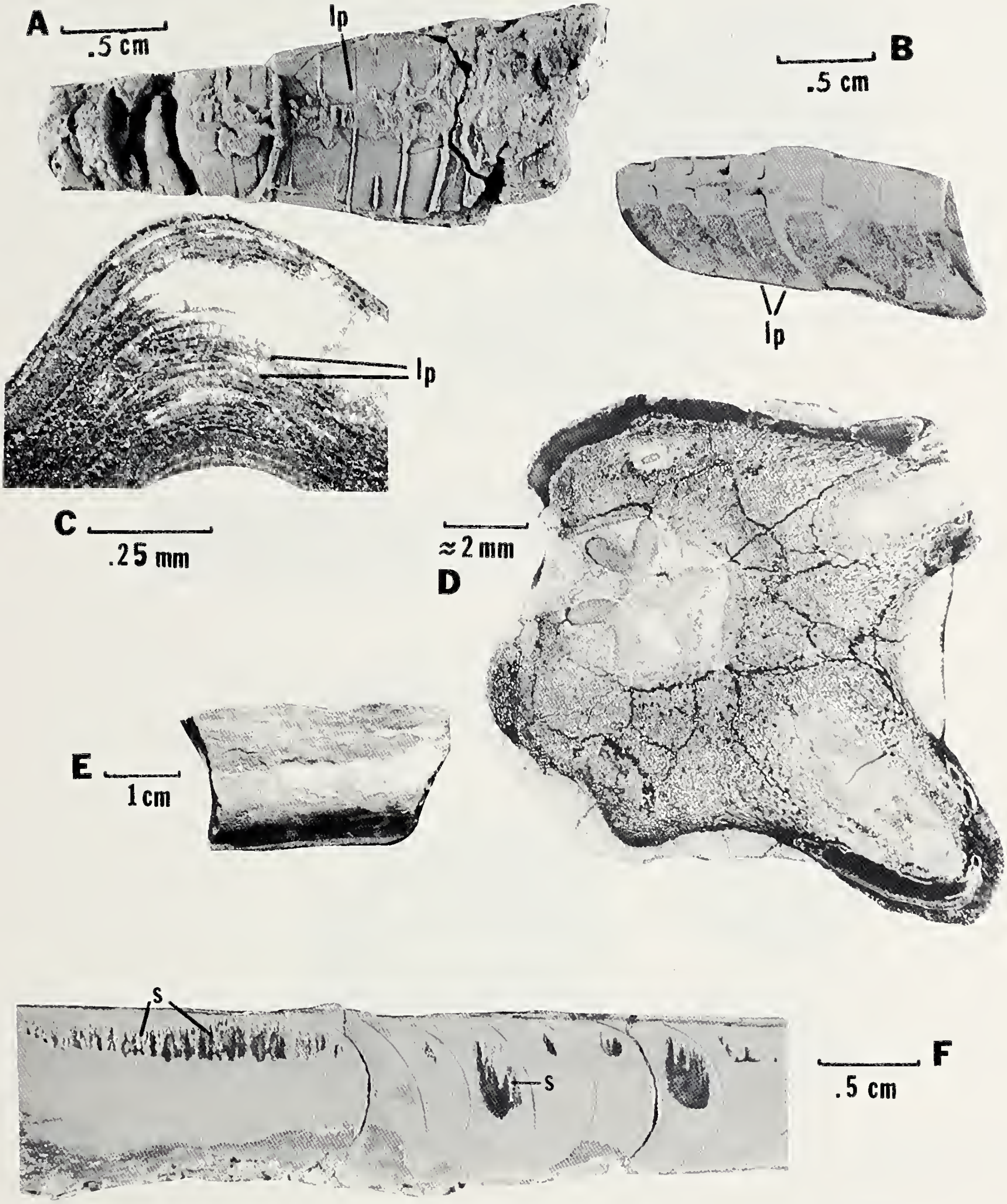


Figure 7: Pyritized wood and nautiloid specimens displaying infilling by forms of early diagenetic pyrite; all from the Windom Shale, Penn-Dixie pyrite bed, locs. 3, 4. Scale indicated by bars.

A. Orthoconic nautiloid *Michelinoceras* sp. Liner pyrite is indicated by "lp" on the shell exterior; RMSC 83. 66. 9.

B. Polished section of *Michelinoceras* sp. Liner pyrite is indicated by "lp"; RMSC 83. 66. 14C.

C. Liner pyrite replacing conchiolin laminae along shell growth surfaces of *Spyroceras* sp. ; RMSC 83. 66. 12.

D., E. Permineralized (by pyrite) wood specimens showing surface (left) and cross-sectioned views (right). Note the excellent cell and tissue structure preservation in the cross-section. Illustration D courtesy of W. Stein.

F. Cross-section of nautiloid showing stalactitic liner pyrite within camerae. Note that these structures act as geopetal indicators; RMSC 83. 66. 14.

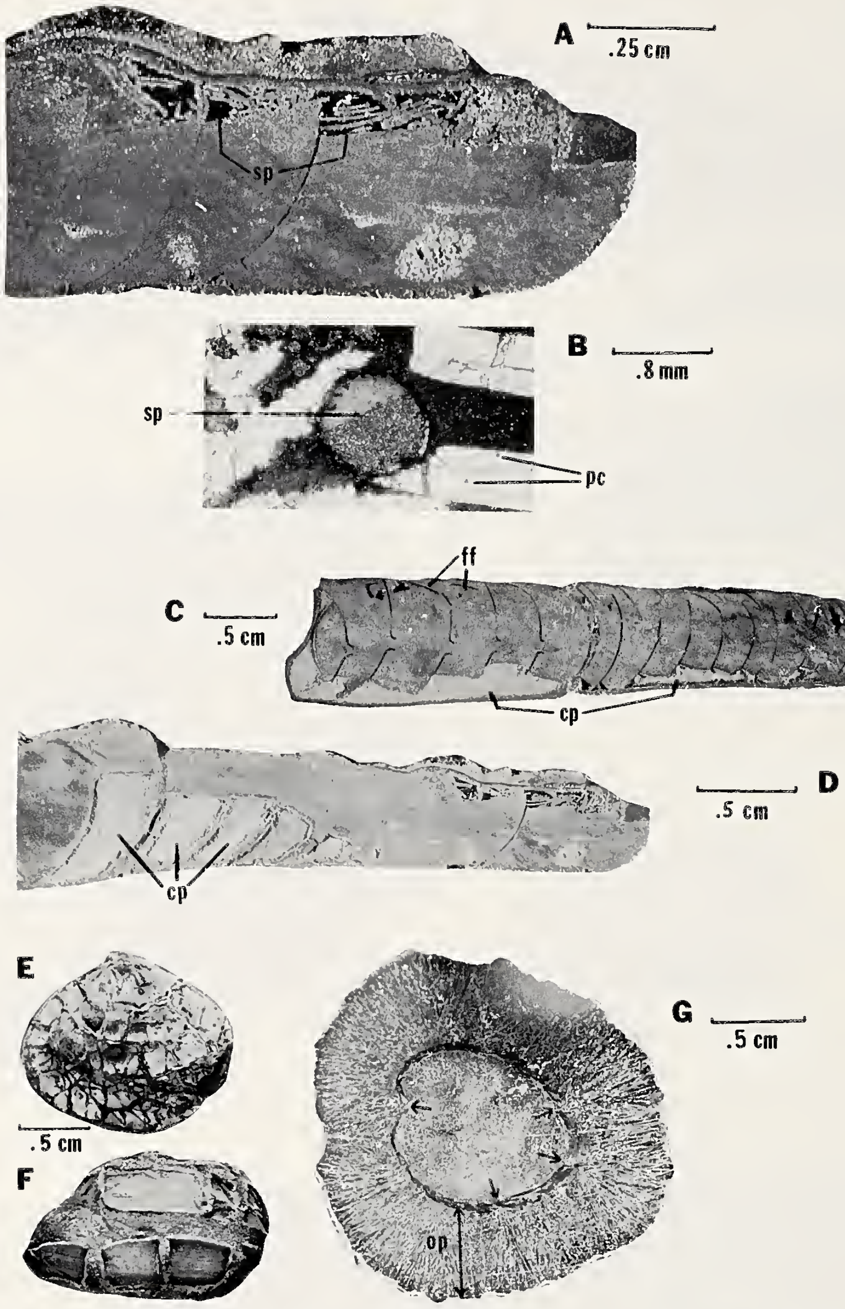


Figure 8: Specimens showing sedimented and calcitic pyrite and later generations of overpyrite. All specimens are Windom Shale, loc. 4, unless otherwise indicated; scales given by bars.

A. Posterior chambers of nautiloid *Michelinoceras* sp. showing level top of sedimented pyrite (sp) indicating upper side of specimen; note crushed fragments of pyritic chamber linings resting on surface of sedimented pyrite; RMSC 83. 66. 14A.

B. Sedimented pyrite (sp) forming a geopetal indicator in an ostracod, all contained within the living chamber of a nautiloid. Also, note the small, isolated pyrite crystals (pc) contained within crystals of barite, which partially fill the cavity; RMSC 83. 66. 12.

C. Cross section of a nautiloid displaying broken fragments of black calcite septae (ff), "floating" in a matrix of sedimented pyrite, demonstrating its unlithified nature during sediment compaction. Calcitic pyrite (cp; enhanced by acid etching) shown flooding adjoining chambers in nautiloid as a result of absence of liner pyrite; RMSC 83. 66. 14B.

D. Calcitic pyrite (again etched) contained within chambers separated by linings of microcrystalline pyrite; RMSC 83. 66. 14A.

E., F. Crack pattern on the exterior of two specimens of *Nuculoidea*; RMSC 83. 66. 2A, B.

G. Cross-section of pyrite nodule nucleated on a nuculid clam. Note the nucleation of the encrusting overpyrite (op) through the expansion cracks in the shell (shown at arrows); Ledyard Shale; Alden bed; loc. 11; RMSC 83. 71. 3.

A further implication of the above is that rapid sedimentation is not always necessary to bury certain shells at the time of death. Therefore, excellent preservation of certain bivalves within pyrite intervals seems to result from an infaunal life habit.

Wood fragments show remarkable preservational differences between the pyritic and non-pyritic assemblages. Remnants of wood occurring outside pyritic interval sediments normally preserve as flattened carbonized impressions. Poor fossilization and compaction destroy cell structure and overall plant morphology. Permineralized wood within Hamilton pyritic intervals retains uncompacted cell walls (Fig. 7D, E), and traces of original organic woody tissue (Stein, 1981).

Similarly, sponge material, when present in non-pyritic assemblages, consists of minute patches of dispersed spicules. Demosponges recovered from Wanakah and Windom pyritic intervals are whole, uncompressed and, occasionally, complete enough to reconstruct nearly entire sponges (Fig. 5G). These examples imply very early permineralization by pyrite.

Trace Fossil Preservation

Threadlike pyritic burrows are present in all Hamilton units and similar in appearance in all facies (Fig. 5A). It, therefore, seems unlikely there is any bias involved in their preservation in different fossil assemblages. Heavily pyritized tubes, however, (Fig. 5F) are only common to pyritic intervals. Tubes are not commonly found except where pyrite is otherwise abundant. Non-preservation, then, may be linked to the lack of conditions appropriate for preservation by pyrite.

Aggregates of fossil shells and steinkerns contained in irregular nodules of pyrite are interpreted as coprolites (Fig. 5B). These structures, when sectioned and polished, reveal an abundance of articulated, disarticulated, and broken valves of brachiopods, bivalves, ostracods and *Styliolina*. Most of the contained shell material is small in size, averaging less than 1 cm in breadth. Packing of shells is tight but there is little order to orientations suggesting that something other than a physical sedimentary process affected the shell masses. The fact that disarticulated valves of originally aragonitic shells are preserved and enclosed with other whole and fragmented shells in a single matrix supports a coprolitic interpretation. Furthermore, these structures greatly resemble nodules from the Fayetteville shale that are also interpreted as coprolitic in origin (Zangerl et al., 1969). In non-pyritic horizons no structures that might be interpreted as coprolites are found except possibly the nuclei of certain carbonate concretions.

Petrology of Fossils and Nodules

Petrographic examination of thin and polished sections of steinkerns and nodules revealed them to be composed predominantly of four minerals plus minor sediment. These are listed below in decreasing order of abundance with notes on variations of each mineral mode (or morphology) of occurrence.

1. Pyrite: Pyrite occurs, at most, in four different forms distinguished by subtle variations in color and crystal habit; these are fine-grained liner or early diagenetic pyrite, sedimented pyrite, calcitic pyrite and macrocrystalline overpyrite.

There has been considerable debate among researchers of Hamilton pyritic assemblages as to whether fossils and nodules are actually pyrite or marcasite (which the crystal form of overpyrite greatly resembles). Izard and Clemency (1967) tested, by x-ray diffraction, samples from fourteen localities of the Alden pyrite and Leicester pyrite in western New York. All were indeed found to be pyrite. Although one, more recent, paper suggested that the bladed crystalline crusts on fossils were actually marcasite (Damagala and Selznick, 1979), x-ray diffraction analyses done for the present paper support Izard and Clemency's findings; both fossil nuclei and "overpyrite" were found to be composed of pyrite. There is probably little, if any, marcasite present.

A. Individual crystals of liner pyrite are primarily cubic or pyritohedral, and appear as a light, grayish brass color in polished sections. This mineral type occurs as thin rinds on and within shells (Fig. 7A, B), patina-like sheets replacing conchiolin along growth surfaces of originally aragonitic shells (Fig. 7C), as a permineralization in wood (Fig. 7D, E), as a stereom infilling in echninoderms and as stalactitic growths in the camerae of goniatites and orthocones (Fig. 7F).

Liner pyrite consistently formed in areas previously occupied by organic tissue. For example, pyritic cameral linings of cephalopods (Figs. 7A, 8A, C, D) replace the area previously occupied by conductive tissue used to secrete cameral deposits (Stenzel, 1964). Pyritic patinas in molluscan shells (Fig. 7C) replace surfaces originally coated with conchiolin (Gregoire, 1964). This close relationship between pyrite and original tissues strongly suggests that pyrite, or at least a precursor of pyrite, began forming before complete tissue decay (see below). This observation also supports the suggestion, made in the previous section, that pyrite formation begins while soft body tissues are still blocking shell apertures.

Occasionally, the shell liner pyrite formed stalactitic extensions, pendant from the roofs of shell chambers (Fig. 7F). Hudson (1978, 1982) attributes pyritic steinkerns to the action of bacterial cultures on mucilage coating actively accreting crystal surfaces.

Lastly, this pyrite type permineralizes wood fragments in relatively unaltered form. Stein (1981) has been able to describe much of the original vascular tissue of *Arachnoxylon kopfii* (Arnold) from fragments found in the upper Windom pyrite and Leicester pyrite beds (Fig. 7D).

B. Sedimented pyrite primarily fills shell voids (especially orthoconic nautiloid chambers) and frequently forms geopetal structures within shell interiors (Fig. 8A, B). It is virtually indistinguishable in crystal form and color from liner pyrite and is thus distinguished only by its mode of occurrence. It apparently settled to the bottoms of chambers at the time of precipitation and filled or partially filled them.

This material is collectively termed sedimented pyrite because it apparently settled to the bottoms of chambers at the time of precipitation and filled chambers, sometimes completely, sometimes incompletely, creating a geopetal fabric. Original horizontal orientations, indicated by geopetal sedimented pyrite, consistently match the orientation of fossil samples collected in the field.

Sedimented pyrite apparently remained unlithified for extended periods and acted as a loose aggregate, often even through compaction. Many samples of partially crushed orthocones display fragments of septal liner pyrite "floating" in or on a sedimented pyrite matrix (Fig. 8A). This could only have occurred if the latter were relatively soft and plastic at the time of compaction and shattering of the pyrite linings. Additional evidence suggests that crystals remained loose even during later diagenesis and were not cemented into a solid pyrite mass until the latest stages of compaction. Figure 8B shows loose pyritic crystals from the interior of a cephalopod apparently enclosed by a later generation of barite crystals. If nodule/steinkern interiors (microenvironments) were indeed so loosely bound, then constituents such as decay gases, mobile chemical species and solutions should have had few barriers to movement during most of the early and late diagenesis period.

C. Calcitic pyrite is framboidal pyrite associated with a ferroan calcite cement and therefore, is considered to be closely associated with sedimented pyrite. The calcitic-pyrite association stands out in thin and polished sections as a dull grayish-brass colored zone, frequently forming a geopetal structure similar to non-calcitic, sedimented pyrite (Fig. 8C, D). When both calcitic and sedimented pyrite co-occur (as cameral fillings, for example), the calcitic pyrite generally is bottom-most, being capped and/or enclosed by the lighter colored liner and/or sedimented pyrite (Fig. 8C, D). Thus, this calcareous pyrite type apparently represents precipitation of ferroan calcite in the interstices of sedimented pyrite crystals or framboids occupying lowest portions of shell chambers. Enclosure of calcitic pyrite suggests that a source of ferroan calcite is internal to the shell, possibly (in part) resulting from a preceding phase of aragonite dissolution.

D. The fourth and last form of pyrite, termed overpyrite (after Hudson, 1978, 1982), consists entirely of large bladed pyrite crystals that partially, or completely, encase fossils, forming nodules and burrow tube linings. The blades always radiate outward, from the nodule nucleus formed by the enclosed steinkern(s), or tube. Overpyrite appears to nucleate from crystals internal to the shell, most often liner pyrite, through points or lines of shell fracture (Fig. 8E-G). Therefore, overpyrite formation must postdate liner pyrite formation and shell fracture. Shell fracture may have been initiated by compaction, but was apparently enhanced by internally generated pressure. Evidence of this may be seen in shells that have fractured along growth lines, with pieces subsequently moving away from one another as internal shell volume apparently increased (Fig. 8E-G). Had the shells simply been fractured by compaction, pieces would be compressed against one another. This expansion may be the result of decomposition gas generation causing gas blowout structures (see Zangerl, 1971), or simply the force of pyrite crystallization. Nevertheless, the result is nucleation and subsequent overpyrite growth.

II. Calcite: three calcite mineral variations occur associated with pyritic fossils and nodules. These variants are distinguished by mineral fabric and results of staining using a potassium ferricyanide-Alizarin Red-S stain (Dickson, 1966).

A. Drusy calcite is the predominant form of the three cements present. It appears light-pink to orange-pink when stained and occurs 1) as crack and fissure fillings (Fig. 9A); 2) as the final cement filling voids spaces of cephalopod camerae or bryozoan zooecia (Fig. 9A); 3) as euhedral "dog tooth" chamber lining, or occasionally as a replacement of aragonitic shell material. Loss of aragonitic shell fabric indicates that this last replacement took place by dissolution and void-refilling by reprecipitation. Cementation (infilling) of compaction-related cracks (of crushed, not expanded shells), further demonstrates that most druse formation occurred very late in the diagenesis of fossils and nodules.

B. Ferroan calcite, the second carbonate type, appears light purple in stained sections. It is most often associated with pore-filling in sedimented pyrite, forming calcitic pyrite (Fig. 8D), but it may also be found occasionally as a skeletal replacement of septa and shell walls in goniatites and nautiloids. Ferroan calcite occupies relatively little volume in most shells, whereas drusy calcite is the predominant void-filler. Ferroan calcite (as calcitic pyrite) is almost invariably found within shell chambers (Fig. 8D, for example). If ferroan calcite is indeed derived from carbonate solutions generated within the shell microenvironment, then dissolution of shell material within the microenvironment (e.g.,septa of cephalopod shells) is a possible source of this calcite type. In addition, this may be a way of determining the timing of pH change within the microenvironment relative to times of formation of other fossil/nodule minerals (see mineralization sequence below).

C. The final calcite type appears to be simply a slightly recrystallized skeletal calcite (e.g., in brachiopods and bryozoans — see Fig. 9A). It appears pink to orange-pink in calcitic shell material when stained and preserves slightly to moderately altered shell-structure. This implies that calcite recrystallization was the result of weak pressure-solution, probably caused by compaction.

III. Barite: Barite always occurs as radiating blades within shell voids. These equant blades originate from individual pyrite crystals, or what appear to be organic nuclei attached to shell walls, or loose within a void space (Fig. 9B). Secondary radiations, or occasionally curved blades, form from original blade tips where individual crystal growth has been obstructed. Otherwise, crystals are predominantly euhedral. Few barite crystals appear to have been fractured by compaction pressure, suggesting that along with drusy calcite, growth was

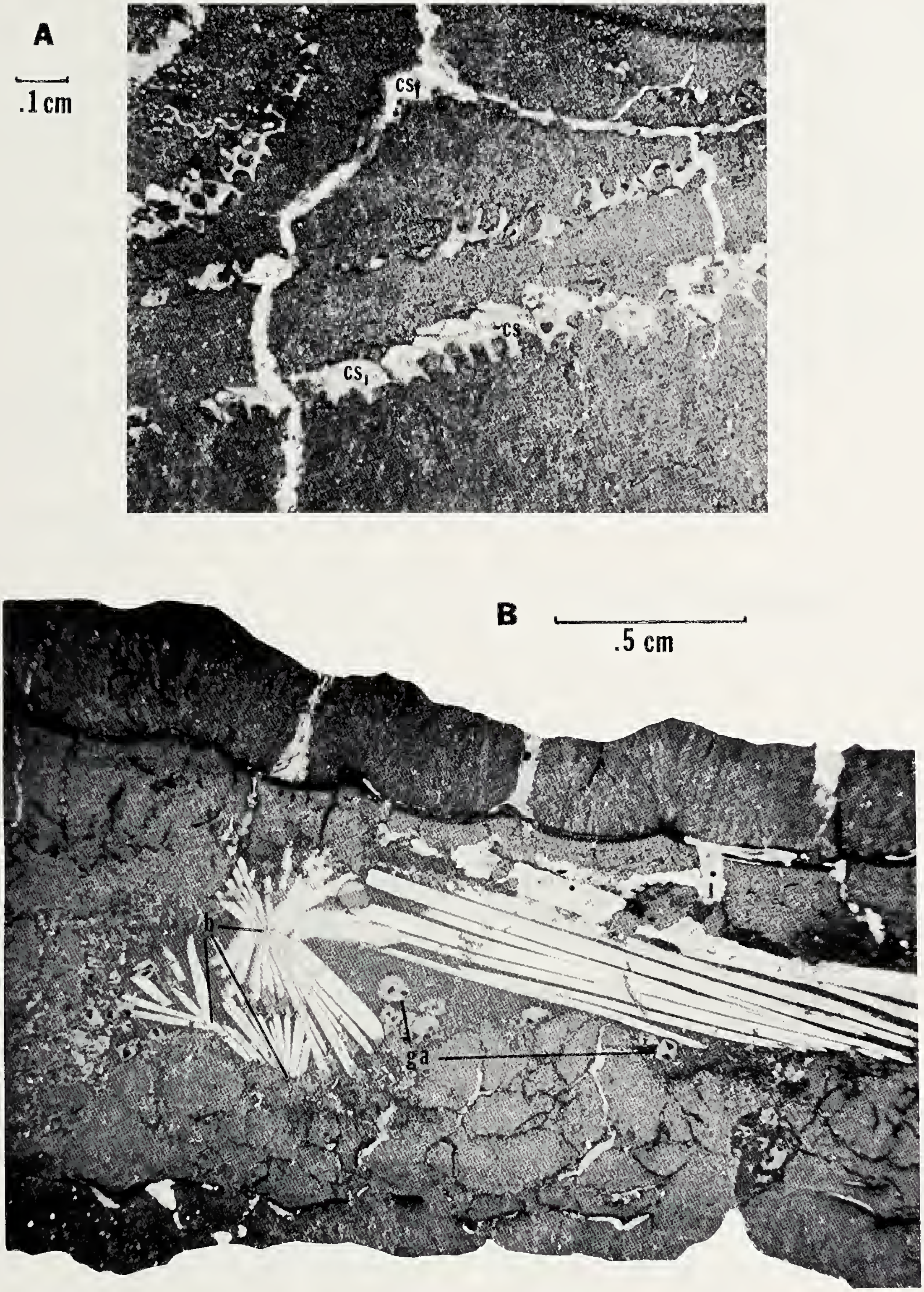


Figure 9: Cross sections of encrusted steinkern nodules showing non-pyrite minerals.
A. Cross-section of bryozoan (*Loxophragma* aff. *L. multitecta*) within pyrite nodule with calcite spar shown filling in cracks (cs1) and replacing original skeletal calcite (cs.); Ledyard Shale, Alden bed; loc. 11; RMSC 83. 70. 3.
B. Interior of *Spyroceras* showing barite blades radiating from points of nucleation (b). Dots on the lower half of the photo are galena crystals with alteration rinds of anglesite (ga). Windom Shale, Penn-Dixie pyrite bed; loc. 4; RMSC 83. 66. 12.

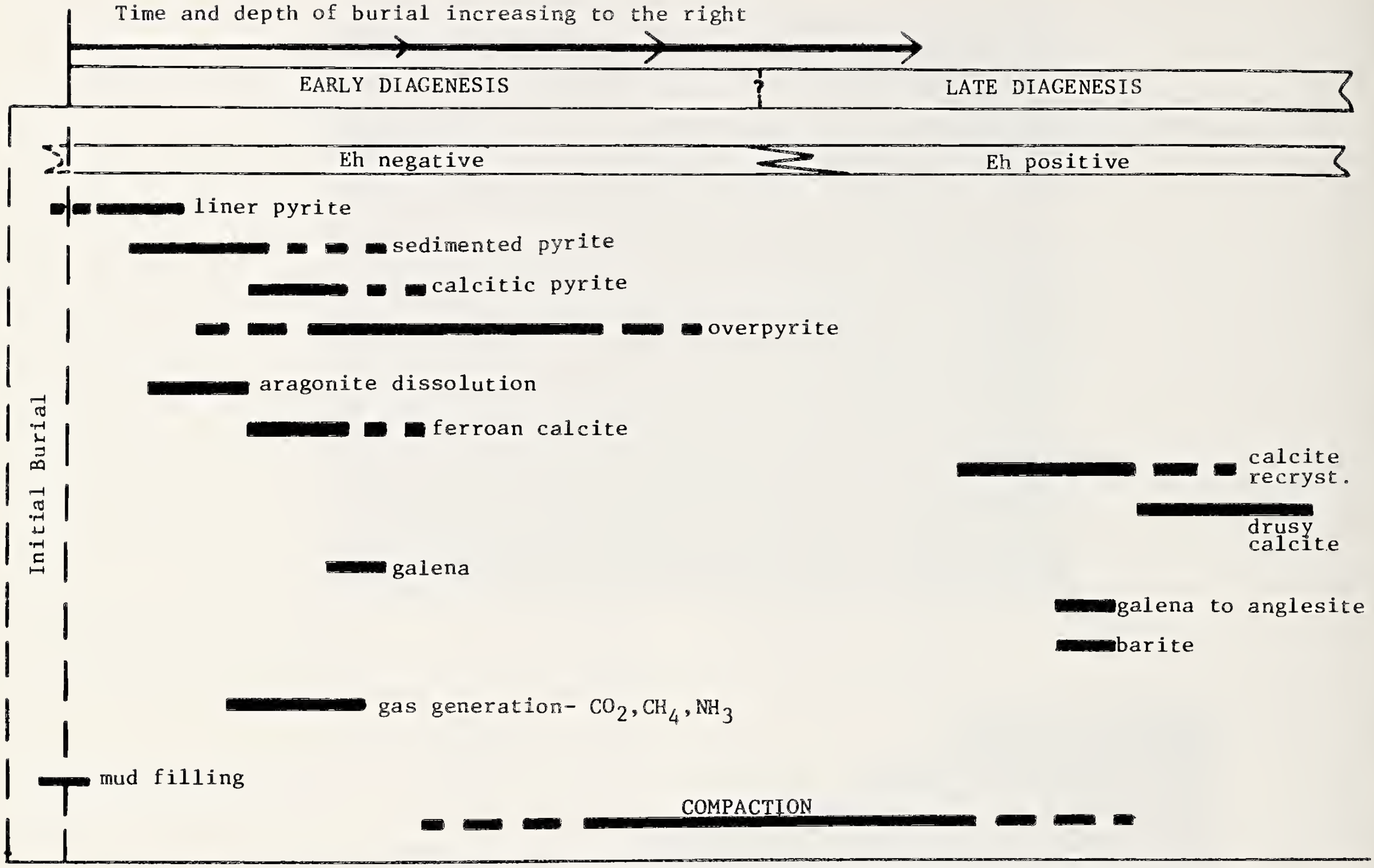


Figure 10: Relative times of formation of minerals included in fossil steinkerns and nodules from Hamilton Group pyritic beds. Horizontal bars indicate duration of mineral precipitation, dissolution or sediment physico-chemical event (e. g., compaction).

a fairly late diagenetic phenomenon, post-dating substantial compaction. Barite was found to be most common in nodules of the Alden pyrite (Ledyard Shale) although minor occurrences were found in Windom pyrite nodules.

IV. Galena: Galena is quite rare, being present primarily, although not exclusively, in steinkerns of the Alden pyrite. Galena crystals are nearly perfect, minute cubes or octahedra, the majority of which have been secondarily altered to display reaction rinds of anglesite (Fig. 9B). These crystals occasionally form a geopetal fabric in the manner of sedimented pyrite. Presence of fractures suggests it is a late forming mineral (i.e.post-compactional). As a reduced mineral species its formation time is probably still quite closely associated with pyrite formation; however, alteration of crystal exteriors to anglesite ($PbSO_4$) is oxidative and may be associated with other similarly oxidized species such as barite that are thought to be late diagenetic (see below).

V. Mud: This is one of the least abundant infilling constituents of pyritic fossils and nodules. When mud is present, it appears as a carbonate-rich clay with finely disseminated pyrite crystals and dark-brown (organic?) inclusions. When present, mud is found filling the living chambers of cephalopods and is rare away from shell apertures. Some examples show flow lines indicating that it probably was still in a viscous state when adjacent mineral growth took place.

Mineralization Sequence

Although many of the minerals identified in this study have previously been documented from Hamilton pyritic assemblages (Fisher, 1951; Kalliokowski, 1966), the sequence and diagenetic timing of mineral formation have never been accurately determined. This has hampered interpretation of diagenesis in the past, making derivation of taphonomic histories largely a matter of conjecture.

Fairly precise ordering of mineral phase formation has been determined in the present study. The interpreted sequence of formation is presented in Figure 10. Mineral formation times have been bracketed relative to one another and, where possible, relative to major sediment physico-chemical events (e.g. burial, compaction, Eh changes).

Because the majority of these minerals are forming in an enclosed space and require points of nucleation, the sequence of formation has generally been interpreted as analogous to geode formation; with crystal growth originating at chamber inner walls and extending progressively inward. Beyond this, cross-cutting relationships and superposition define the order of formation. Thus liner pyrite is the first formed mineral in most examples, followed by sedimented pyrite, calcitic pyrite, and so on.

A number of general features and implications of Figure 10 require emphasis. First, the sequence of formation is derived from many specimens, not all of which contain precisely the same mineral phases or exhibit an identical mineralization sequence. Thus Figure 10 is, by necessity, a composite sequence and times of crystallization must overlap. Generally, those mineral phases or events that are most commonly seen in fossils and nodules are placed at the top of the chart and least common ones at the bottom. The phragmocones of orthocones and goniatites have been most useful in deriving this sequence and, therefore are most often referred to in the following text and figures. However, the same mineralization features apply to all shell chambers.

Overlap of formation times results because, in many examples, the formation of one mineral phase appears to have been partly controlled by a preceding mineral phase. For example, replacement of aragonite by a secondary calcite in molluscan shells seems to depend upon nucleation and growth of liner and overpyrite. Non-encrusted steinkerns (Fig. 8A-F) result when no overpyrite nucleation occurs and subsequent dissolution of the aragonitic shell material exposes an internal pyritic mold. Conversely, complete shell replacement (Fig. 8G) results when the fossil has been covered entirely by overpyrite, before total shell dissolution. Thus later diagenetic carbonates could fill the void left by aragonitic shell dissolution. This void would have been enclosed by liner and overpyrite.

A second example of mineralization overlap involves the formation of calcitic pyrite. Occasionally calcitic pyrite will occupy the bottom portions of several adjacent cephalopod chambers, unseparated by liner pyrite encrusted on septa (Fig. 8C). If the formation times of liner and sedimented pyrite overlap (Fig. 10), it is possible for them to compete for space within the microenvironment. This allows sedimented pyrite to be deposited against the bottoms of septa before liner pyrite completely lines a chamber. Subsequent introduction of ferroan calcite will equally fill adjacent chambers not separated by a liner pyrite and provide a mechanism to account for various distributions of ferroan calcite, at least within chambered shells. In summary, the phases of mineral formation in any single steinkern may be interpreted, but by comparing a number of steinkerns it is clear that a certain degree of interdependence exists, governing mineral formation.

A second implication of Figure 10 is that compaction seems to have affected, to a minor degree, most of the mineral species to the left of the figure but those phases to the right of the figure remain relatively unaffected. The compaction event also allows convenient division of those minerals that form in predominantly negative Eh environments (left of Fig. 10) from those formed in positive Eh environments (right of Fig. 10). Corresponding areas of negative and positive Eh have been plotted on the diagram to reflect this. The Eh line reflects primarily overall sediment chemical conditions and not localized conditions of microenvironments. For example, overpyrite growth is the first indication that overall (macroenvironmental) sediment conditions are Eh negative, a condition necessary to allow pyrite growth on shell exteriors. All pyrite growth prior to this has been internal to shells (microenvironmental), and we infer reducing, sulfidic conditions within the shell microenvironment. It is possible that overall sediment conditions were slightly oxidizing, as a result of bioturbation mixing the upper few centimeters of sediment while the first internal pyrite was forming. It is difficult to estimate to what depth below the sediment-water interface oxidizing conditions extend, although it must be no greater than the depth of bioturbation. Therefore, the area of negative Eh (left side of Fig. 10) is drawn to reflect overlapping micro- and macroenvironment conditions.

Gravitational compaction is generally the effect of load pressure on sediment volume with increasing overburden resulting in decreasing pore space. Although a wide range of porosity values at depth have been documented (Zangerl and others, 1969), an average estimate for clay sediments reaching a plastic limit (about 45% porosity — Weller, 1959) is 10 m or more. Kepferle (1977, p. 44) argues for significant compaction of muds at depths as shallow as 1 m. Thus, the bar showing first effects of compaction on Figure 10 may correspond to a depth of 1 to 10 m.

Figure 10 is intended to reflect relative times of formation and in this sense the sequence shown has good applicability to derivation of depositional and geochemical models (see geochemistry section below).

GEOCHEMICAL MODELS

Iron within Pyritic Beds

Previous authors (Loomis, 1903; Fisher, 1951) have called upon extraordinary sources such as "iron water" to explain the formation of Devonian pyritic beds but significant evidence supports early diagenetic formation of pyrite within certain fossil forms by normal sediment-geochemical processes (Brown, 1966; Kalliokowski, 1966; Hudson, 1978, 1982; Clarke and Lutz, 1980; Reaves, 1984).

The amount and distribution of total iron (Fe^{+2} and Fe^{+3} were determined for the Alden pyrite of the Ledyard Shale by wet chemical analyses and by measurement of total pyrite mass for a given volume of sediment; three observations resulted.

First, bulk rock levels of iron for the Alden pyrite average between 4% and 6% by dry weight. McCollum (1980) similarly determined iron levels for the total Ledyard Shale Member to range from 4% – 7%. "Average shale" iron contents are within this range (Krauskopf, 1967). Therefore, claims of iron being unusually enriched in Hamilton waters seem unfounded and pyrite formation occurred in a "normal" sediment.

Second, by determining the total mass of pyrite (fossils and nodules) collected from a given mass of sediment, it was possible to estimate the pyrite component of iron in Ledyard sediments. Table 2 shows the amount of pyrite derived from a given sample and the corresponding mass of iron contained in this quantity. Also shown is the percentage of this pyrite iron as compared to the total volume of sediment sampled. Note that these latter figures range from .02% – 0.20% of Fe per unit sediment. These calculations are inevitably low due to the impossibility of collecting all pyrite (especially disseminated framboids) from within a sample, but it is felt that sampling was careful enough that these figures are at most an order of magnitude below the correct amount of pyrite iron for the given sediment volumes. The estimates demonstrate that pyrite did not act as an iron "sink" within Hamilton pyritic beds; that is, no unusual amounts of iron are being used to fossilize the forms observed. Indeed, of the total iron that was available in these shales only a very small percentage seems to have been used to form pyrite.

Geochemistry and Early Diagenesis

Taphonomic history of Hamilton Group pyrite beds may be explained by a synthesized geochemical model that best fits qualitative parameters derived from the biostratinomic, sedimentologic and petrologic evidence described herein. A modified "high reactive iron" model (Berner, 1969) of iron and sulfur migration in unconsolidated sediments is used to explain early diagenetic pyrite formation in Hamilton intervals. This model requires:

1. A large, but unevenly distributed quantity of organic matter in the original sediment.
2. A large amount (approximately 0.5 – 0.8% by dry weight) of reactive iron (as distinguished from total iron, which is about 4 – 6%).
3. A low sedimentation rate, coupled with episodic sedimentation events to bury organisms.

Berner's (1969, 1971) original model calls for the existence of reducing conditions from the site of decomposition on down into the sediment (see Berner, 1971, Fig. 6 – 8). The present model is modified in that, as determined by stratigraphic and biostratinomic evidence, the redox potential discontinuity probably lay approximately 3 – 5 cm below the sediment-water interface. Therefore, decaying organisms experienced reducing microenvironments in shell interiors, as well as slightly oxidizing macroenvironmental conditions from 0 – 5 cm below sediment-water interface, and reducing macroenvironmental conditions at 5 cm and deeper below the sediment-water interface. This is shown schematically in Figure 11.

Before discussion of Berner's model, we must consider how well parameters of Hamilton pyrite beds fit the three requirements outlined above.

First, organic carbon (C_{org}), at least in the Alden pyrite, ranges between 0.4 and 0.8% (McCollum, 1980). Based on correlations between rock color and organic carbon content (Potter and others, 1980, Fig. 1.25), other pyrite beds in the upper Hamilton fall within this same range. This is a moderate to low level of C_{org} for shales in general with the most organic rich containing roughly 1.0 – 5.0% of organic carbon (Potter and others, 1980). Furthermore, C_{org} levels indicated for Hamilton sediments are bulk rock levels of organic carbon. Field evidence suggests that distribution of organics at the time of deposition was much more patchy than both analyses indicate, with concentration of material in localized shell patches, pavements, or thin layers separated by a few to tens of centimeters of barren sediment. Therefore, periodic accumulation of organisms in localized areas seems to have produced a relative concentration of organics over moderate to low bulk sediment C_{org} levels, thus satisfying requirement 1.

Second, as indicated previously, the amount of iron incorporated into pyrite from measured intervals converts to an original range of .02 – 0.20% reactive Fe by dry weight of sediment (Table 2). This amount of iron is only slightly lower (0.3%) than the amount of reactive iron required by Berner's high reactive iron model. The impossibility of recovering finely disseminated pyrite from each interval would account for these low estimates. Again, if we use the criteria of Potter et al., (1980, Fig.1.25) and use the correlation between color and mole% Fe^{+2}/Fe total, then pyrite intervals fall in the correct range of 0.6 – 0.8% reactive Fe by dry weight, and the second requirement is satisfied.

Finally, there is indirect evidence for slow background sedimentation rates and there is direct evidence for episodic rapid sedimentation causing burial and good preservation of selected communities. Presence of certain organisms that require non-turbid water (sponges, blastoids, bryozoans) to live implies that, at times, sedimentation rates were at a minimum. Therefore, the third requirement of Berner's model also is met.

Mechanisms of Sedimentary Pyrite Formation

Sedimentary pyrite forms in low Eh, anoxic settings. These conditions may be widespread, occurring in dark muds (below the sediment-water interface) of tidal flats or stagnant basins, or in localized reducing microenvironments in otherwise oxygenated sediments.

Pyrite is the eventual product of ferrous iron and sulfide combining through the bacterially mediated decomposition of organic matter. In modern marine settings the bacteria *Desulfovibrio* and *Desulfotomaculum* reduce sea water SO_4^{-1} to

Locality	stratigraphic interval	shale sample weight ($x10^3$kg)	weight of pyrite in sample (kg)	Fe_{pyrite} (=46.5% pyrite wt.)	estimated $Fe_{reactive}$ ($=Fe_{pyrite} \div$ shale wt.)	$\%Fe_{total}$ contained in pyrite sampled ($=Fe_{pyrite} \div Fe_{total}$)
Alden pyrite Locality 15	interval F/G	3.65	16.42	7.65	.21%	5.2%
	interval G/H	3.77	2.57	1.19	.03%	7.4%
Alden pyrite Locality 11	interval E/F	1.14	0.65	0.30	.03%	0.7%
	interval F/G	1.21	0.57	0.27	.02%	0.6%
	interval G/H	0.97	0.59	0.28	.03%	0.7%
	interval H/I	0.69	0.50	0.23	.03%	0.8%
	interval I/J	0.92	2.03	0.95	.10%	3.0%
	interval J/K	1.24	0.49	0.23	.02%	0.5%

Table 2: Volume data (in kg and % sampled) is presented here for pyrite collected from Alden pyrite localities 11 and 15. All macroscopic pyrite was collected from sample volumes of shale in roughly 0. 1 to 0. 3 m^3 intervals, Fe pyrite was calculated using the mole % of iron in FeS_2 (= 46. 5%). Fereactive gives an approximation of the percentage of iron from a shale sample volume "used" to form the pyrite derived from each sample. Percentages given in the last column reflect the percentage of shale total iron contained in the sampled pyrite, assuming the matrix shale is 4% Fe by dry weight.

H_2S during the metabolic breakdown of organic matter. This process has been shown to account for the bulk of the sulfide introduced to sediments in modern environments (Berner, 1969; Goldhaber and Kaplan, 1974). Reactive iron, introduced to sediment in the form of iron hydroxides (Berner, 1964; Rickard, 1975), limonitic coatings on clastic grains or adsorbed on clays (Krauskopf, 1967), becomes reduced under anoxic conditions.

Pyrite may form from ferrous iron and sulfide by two general pathways. Most often, Fe^{+2} and H_2S first react to form various iron monosulfides (mackinawite, griegite or amorphous FeS — see Berner, 1964) which are then secondarily converted to pyrite by the addition of elemental sulfur to the iron monosulfide (Berner, 1969, 1971; Rickard, 1975). This solid state reaction may require several days to months to complete (Berner, 1969, 1970). Secondly, the direct formation of pyrite at lower pH (5.0 – 6.5) and pH_2S may occur through various heterogeneous reactions (Rickard, 1975), and yield products in a matter of one to several days (Howarth, 1979). The latter process occurs under exceptional circumstances and is not likely to have been important in generating Hamilton pyrites.

Processes of Iron and Sulfur Concentration

Pyrite constituent migration in the modified high reactive iron model are schematically represented in Figure 11 and are discussed below.

Concentration of organics in widely spaced, discrete patches upon or within sediments initially allows immobilization of reduced sulfur (HS^{-1}, H_2S) at, or proximal to, sites of first organic decomposition. This is based on the hypothesis that the first-generated sulfide resulting from bacterial reduction would rapidly combine with reactive iron from immediately adjacent sediments to generate some form of iron monosulfide (e.g.mackinawite, griegite, etc.)or pyrite directly. It is essential at the outset to confine sulfide migration to the immediate area of organic decomposition by this initial reaction. Escape of H_2S into surrounding sediments (such as in the "low reactive iron" model of Berner, 1969) would trap reactive iron as a monosulfide away from the decomposition site and exclude it from further diffusion toward the area of fossil mineralization. However, with confinement of H_2S at the microenvironment site, Fe^{+2} distal to the decomposing matter remains as a mobile chemical species. Both situations described above, the "high" and "low" reactive iron models, have been qualitatively demonstrated in the laboratory (Berner, 1969), and thus, this initial mechanism seems workable.

Initial combination of ferrous iron and sulfide, and subsequent formation of pyrite within shells, may proceed by either solid state or heterogeneous reactions, and depends primarily on initial pH and pH_2S. Even if initial chemical conditions in the microenvironment favor the formation of mackinawite and griegite, such iron monosulfides are metastable in the presence of crystalline pyrite and will spontaneously convert to pyrite (Berner, 1970; Rickard, 1975). Therefore, first formation of FeS_2 by either reaction mechanism will enhance further pyritization from any pre-existing monosulfides.

After initial uptake of reactive iron immediately adjacent to the decomposition site, further concentration of Fe^{+2} depends on diffusion through the sediment column and secondarily on physical movement of iron-bearing sediments in the area near the site of decomposition. Figure 11 shows that Fe^{+2} diffusion into a reducing microenvironment may be partially driven by the outward diffusion of other products of bacterial decomposition, such as methane and ammonia. These species are not as reactive as sulfides toward iron (Berner, 1971), and their effect is to reduce ferric iron (from ferric hydroxide compounds) to Fe^{+2} aq. and thereby free iron to diffuse toward the pyritization area.

Physical movement of available iron from deep sediments (5 – 10 cm) might occur biogenically due to processing of detritus by infauna. "Mining" of deep sediments by the Recent tube dwelling polychaete *Clymenella* has been shown to double the amount of iron in the upper few centimeters of sediment over that which would be present in uninhabited sediment (Aller, 1978). *Clymenella* is described as a "conveyor belt" organism that processes detritus at depth and deposits digested sediments at the sediment-water interface. A necessary outcome of this process is the concentration of reactive iron near the decomposition site. Thus, constant reworking of sediments by tube dwellers and other infauna may also be a factor in speeding diffusion rates (Rickard, 1975).

By application of the modified high reactive iron model, we may, at least qualitatively, describe mechanisms that concentrate and combine iron and sulfur at sites of organic decomposition to form pyritic fossil steinkerns, nodules and crusts. Additionally, it is possible to infer the effects of each of the three diagenetic subenvironments of the model (reducing microenvironments, and oxidizing and reducing macroenvironments in Fig. 11), by relating mineral types in pyritic nodules and steinkerns to each of three subenvironments of the formation.

First, early diagenetic liner pyrite and at least some of the sedimented pyrite probably are products of microenvironment processes taking place within the upper few centimeters of sediment. Initial pyritization may occur by any of the mechanisms already discussed. Subsequent ferroan calcite precipitation in the microenvironment may be caused by gradual elevation of alkalinity resulting from generation of HS^{-1} and H_2S and to a lesser extent generation of NH_4 and NH_3. Depending on relative amounts of each of these compounds, final pH will range between 7 and 9 in the microenvironment thereby allowing carbonate precipitation.

Second, aragonite dissolution from shell exteriors may also be accomplished in the upper few centimeters of sediment but as a result of the oxidizing macroenvironment. Dissolution is a product of slight lowering of pH that may result from partial decomposition of organics by aerobic bacteria. Carbon dioxide generation from decomposition forms a weak carbonic

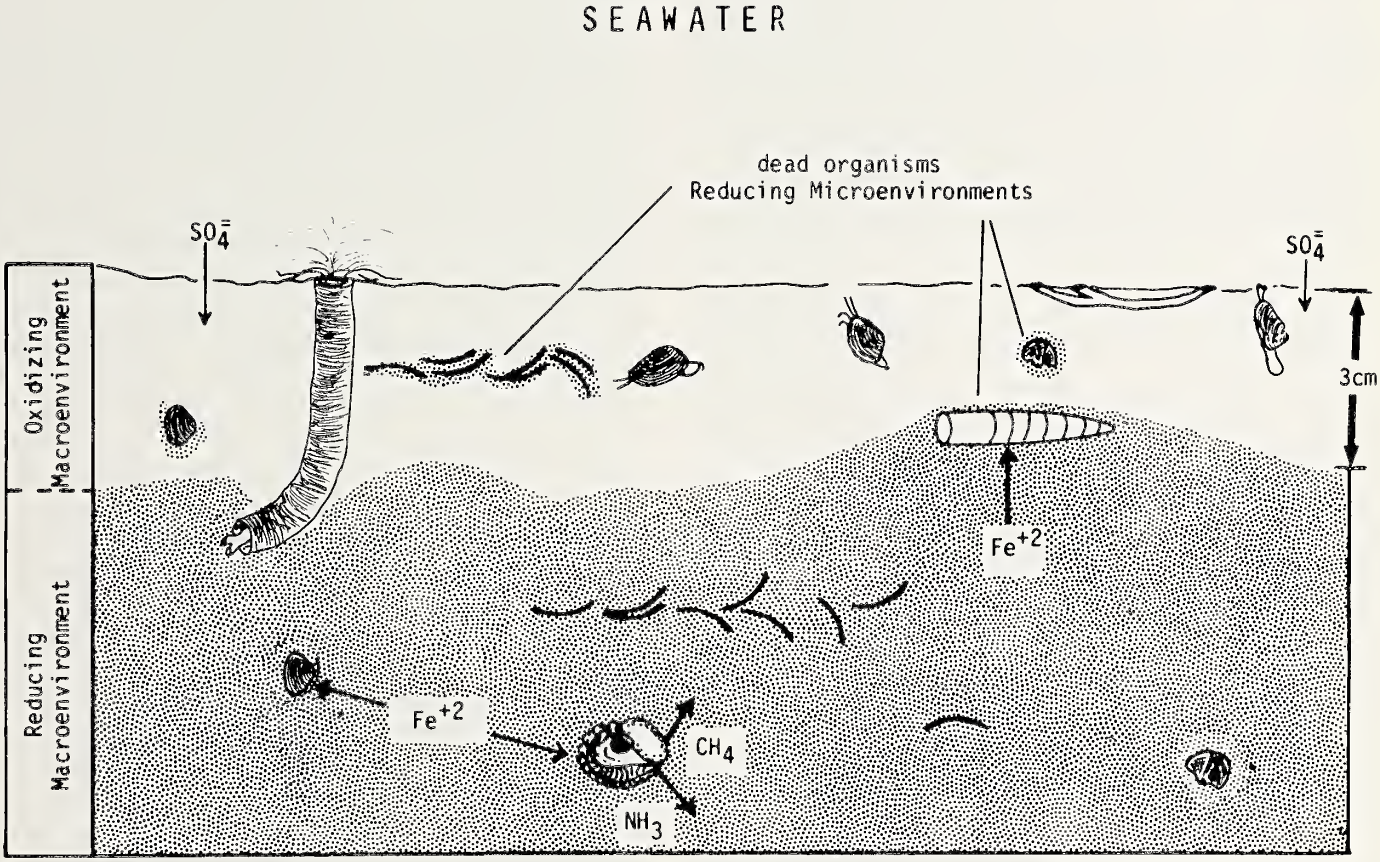

SEAWATER
dead organisms
Reducing Microenvironments
$SO_4^=$
$SO_4^=$
3cm
Oxidizing Macroenvironment
Reducing Macroenvironment
Fe^{+2}
Fe^{+2}
CH_4
NH_3

Figure 11: Movement of chemical constituents between and within three substrate environments of upper Hamilton Group pyrite subfacies. See text of the Geochemistry section for discussion of chemical reactions allowing constituent movement.

acid (see Richards, 1965; Berner, 1971). Intermittent oxidation of iron monosulfides would also produce sulfuric acid in the upper sediment (Reaves, 1984, p.321). The acidity produced is presumed to be sufficient to cause aragonite dissolution but leave calcite relatively unaffected. Varying thicknesses of this oxidizing macroenvironment may account for variance in the amount of aragonitic shell material preserved (through recrystallization or replacement). Residence time of a shell in a thin oxidizing macroenvironment (shallow RPD) would be short, resulting in minimal shell loss. Conversely, in a thick oxidizing macroenvironmental zone (deep RPD) the shell would be subject to acidic conditions comparatively longer, decreasing the chances for shell preservation.

Third, the passage of shells into the realm of the reducing macroenvironment is signalled by growth of bladed overpyrite. Reducing macroenvironment conditions must be present to allow accretion of reduced minerals on shell exteriors.

Recently, Berner (1981) proposed a new sediment geochemical classification system. Application of this classification to Hamilton pyritic bed environments would result in the labelling seen in Figure 11. The upper few cm (0 – 5 cm) of mud, the oxidizing macroenvironment, would most likely fall into Berner's "oxic" category, but dissolved oxygen levels probably would be only 0.1 – 1.0 ml/l (based on Rhoads and Morse, 1971). Within the oxic environment localized sites of decomposition (shell interiors) produce microenvironments that might be classified as "sulfidic" depending on actual pH_2S values. Finally, below 5 cm to an undefinable depth, anoxic "non-sulfidic" conditions would exist, again depending on actual Hamilton sediment pH_2S values.

DEPOSITIONAL MODEL FOR HAMILTON PYRITIC BEDS

Recently, Reaves (1984, p. 318) has outlined the sedimentary conditions requisite for pyrite aggregate formation based on field and experimental study of Recent sediments. Features which maximize formation of pyrite in local aggregations include: 1) well oxygenated bottom water; 2) low average sedimentation rates coupled with 3)episodic burial of the bottom by resuspended muds; 4) autochthonous benthic fauna; 5) warm climate and 6) absence of large seasonal fluctuations of sediment oxygenation and temperature. These factors promote burial of localized organic nuclei in sediments that are generally poor in metabolizable organics. All of these conditions appear to have existed in the depositional environments of Hamilton pyrites.

Upper Hamilton Group pyritic beds probably developed in the upper dysaerobic zone which supported a diverse fauna composed of both infaunal and epifaunal forms. Dissolved 0_2 levels were improved over the lower two zones and oxic conditions probably extended to a shallow level within the sediment. The upper dysaerobic zone substrate was locally thixotropic as a result of constant reworking by a diverse infauna. Sedimentation was slow and bottom waters were relatively clear most of the time. Sudden environmental changes, such as increased turbidity or onset of anoxia, resulted in preservation of rare fossil assemblages, such as blastoid and sponge horizons. Factors of the upper dysaerobic zone that probably greatly influenced early diagenetic pyrite development were:

1. Unevenly distributed, locally abundant communities of organisms (and thus organic matter) together with moderate to low background organic matter levels.
2. Dissolved 0_2 levels high enough to support a fauna but low enough to preclude substantial aerobic decomposition of organics in the upper few centimeters of sediment (thus allowing predominantly anaerobic micro- and macroenvironmental decomposition).
3. Abundance of reactive iron in fine-grained sediments.
4. An active infauna that co-mixed sediments, thereby enhancing diffusion rates of pyrite constituents.
5. Generally low background sedimentation punctuated by episodes of rapid decomposition that suddenly buried organic matter.

Determination of taphonomic histories of fossil forms allows further interpretation of Hamilton Group depositional environments. Biostratinomic evidence supports the conclusion that periodic rapid influx of sediment caused burial and preservation of delicate fossil forms that otherwise existed under conditions of low to moderate background sedimentation (cf. Brett and others, this volume). Diagenetic and biostratinomic evidence implies further that in a stratified basin model (after Rhoads and Morse, 1971) the redox potential discontinuity lay at or near the sediment-water interface in the anaerobic zone, and gradually descended below the sediment-water interface through the dysaerobic zone and still deeper into the aerobic zone (Fig. 12). Enrolled trilobite and nodule horizons suggest periodic elevation of this boundary, possibly caused by storm generated changes in bottom currents, salinity and/or organic influx.

Sedimentologic, stratigraphic and taphonomic evidence presented herein for pyritic bed deposition, and the demonstrated workability of the high reactive iron model to form substantial early diagenetic pyrite in the laboratory (Berner, 1969; Rickard, 1975; Howarth, 1979) strongly contradicts any claim that Hamilton pyritic intervals are common only to anoxic black shale facies, or that they are the product of extraordinary water or sediment chemistry, as suggested by previous workers (Loomis, 1903; Fisher, 1951). Pyritic fossils and nodules were most probably formed at a specific level along shore to basin environmental gradients in a density stratified basin by normal sediment physico-chemical processes. Furthermore, if the litho- and biofacies belts that formed in these gradients generally trend parallel to the present Hamilton outcrop belt (east-west), then we may explain the relative uniformity of rock thickness, and lithologic and paleontologic composition across the Hamilton outcrop belt in western New York. Localized minor variation in faunal diversity, sediment type and unit thickness within pyritic intervals may then be a result of north-south deflection of facies belts relative to the present outcrop belt.

Reaves (1984) suggests that black to gray shale transitions, such as those observed in the Hamilton Group, may actually reflect differences in the quantity of metabolizable organic matter supplied to sediments overlain entirely by oxygenated water, hence obviating postulation of basinal oxygen stratification. However, we do not accept this explanation for Hamilton Group dark shale facies because of the constraints of facies geometry. Black shales almost invariably occur in a central zone (Appalachian Basin center) and grade laterally into concentric belts of gray mudstone, calcareous beds or siltstone. Reaves' model would require that higher amounts of metabolizable organic matter be supplied to the central zone leading to black shale development. However, this is unlikely; nearshore areas would be more likely to receive higher inputs of nutrients and metabolizable organics. Also the reduced diversity or complete absence of benthic fossils in black shale facies indicate that water overlying the sediment was poorly oxygenated or anaerobic. Instead, the Hamilton facies geometry closely resembles the pattern seen in Recent stratified basins in the California continental borderland described by Savrda and others (1984). The stratified basin zonation of Rhoads and Morse (1971) was first developed to describe modern environments on the continental shelves of North America; it has been modified slightly by the discovery of small calcareous, epifaunal organisms in environments with 0_2 levels as low as 0.3 ml/l (Thompson and others, 1985). Byers (1974, 1977) first adapted this zonation to Appalachian basin shales of New York State and Pennsylvania and adequately summed up evidence for stratification of the Appalachian basin during Upper Devonian time. Evidence gathered for this study supports the conclusions of others (McCollum, 1980; Baird and Brett, 1981; Ettensohn and Barron, 1981) that stratification of the Appalachian Basin began in the Middle Devonian; perhaps even before Hamilton Group sediment deposition. Stratified basin deposition may be quite helpful in explaining the formation of a number of other pyritic fossil occurrences (see, for example, Kammer, 1981). Of course this type of model is not a panacea for all pyrite-rich sedimentary deposits, or similar stratification occurrences within shale sequences. It is hoped, however, that application of the same type of observational techniques that aided the progress of this study, and use of actualistic models to explain pyritic bed development, will permit closer evaluation of other pyritic assemblages. Only through these means will the proper interpretation of shale depositional environments be accomplished.

ACKNOWLEDGMENTS

We are grateful to G.Richard Holdren for his assistance with geochemical analyses and for many constructive discussions during this study. Charles Clemency aided with x-ray diffraction studies of pyrite, Michael Savarese, David Lukasik and Eric Fisherman assisted with fieldwork, Robert Eaton prepared many of the photographs and Margrit Gardner patiently typed several drafts of the manuscript.

Various versions of this paper were reviewed by Thomas Kammer, Chris Reaves, G. Richard Holdren, and Curt Teichert, who provided numerous criticisms and suggestions that greatly improved the final product.

Specimens figured in this paper are reposited in the collections of the Rochester Museum and Science Center (RMSC). Finally, we appreciate the financial assistance provided by Rochester Museum and Science Center, a Geological Society of America research grant, the donors to the Petroleum Research Fund, American Chemical Society and National Science Foundation Grant EAR-8313103.

REFERENCES CITED

Aller, R.C. 1978. Experimental studies of changes produced by deposit feeders on pore-water, sediments and overlying water chemistry. Amer. J. Sci. 278:1185-1234.

_________ 1980. Relationship of tube-dwelling benthos with sediment and overlying water chemistry. *In* Tenore, K.R. and Coull, B.C., eds., Marine Benthic Dynamics, Univ. of South Carolina Press, Columbia, S.C., p. 285-308.

Babcock, L. E. 1982. Original and diagenetic color patterns in two phacopid trilobites from the Devonian New York. Proc. 3rd North Amer. Paleont. Conv. 1:17-22.

_________ and Speyer, S.E. 1983. Coiling in two Devonian phacopine trilobites: a response to anaerobiosis. Geol. Soc. Amer. Abstr. with Programs 15(2):171.

Baird, G.C. and Brett, C.E. 1981. Submarine discontinuities and sedimentary condensation in the Hamilton Group (Middle Devonian): examination of marine shelf and paleoslope deposits in the Cayuga Valley region. *In* New York State Geol. Assn. Guidebook, 53rd Ann. Mtg., SUNY at Binghamton, p. 115-145.

Beerbower, J.R. 1981. A paleomicroecologic study of deposition and early diagenesis in a Mid-Devonian marine shale. Geol. Soc. Amer. Abstr. with Programs 13:406.

_________, Bray, R.G., Buehler, E.J., and Jordan, F.W. 1969. A "paleomicroecologic" study of Mid-Devonian marine assemblages (Hamilton Group) in western New York. *In* Guidebook of the North Amer. Paleont. Conv., Trip 5, Paper C, p. 4877.

Berner, R.A. 1964. Iron sulfides formed from aqueous solution at low temperature and atmospheric pressure. J. Geol. 72:293-306.

_________ 1969. Migration of iron and sulfur within anaerobic sediments during early diagenesis. Amer. J. Sci. 268:19-42.

_________ 1970. Sedimentary pyrite formation. Amer. J. Sci. 269:1-23.

_________ 1971. Principles of Chemical Sedimentology. McGrawHill, Inc., New York, 240 p.

_________ 1981. A new geochemical classification of sedimentary environments. J. Sediment. Petrol. 51:359-365.

Bowen, Z.P., McAlester, A.L., and Rhoads, D.C. 1974. Marine benthic communities of the Sonyea Group (Upper Devonian) of New York. Lethaia 7:183-220.

Bray, R.G. and Beerbower, J.R. 1969. Paleoecologic analysis of some Middle Devonian fossil aggregations. Amer. Assn. Petroleum Geol. Bull. 53:709.

Brenchley, P.J. and Newall, G. 1970. Flume experiments on the orientation and transport of models and shell valves. Palaeogeog., Palaeoclimatol., Palaeocol. 7:185-200.

Brett, C.E. 1974. Biostratigraphy and paleoecology of the Windom Shale Member (Moscow Fm.) in Erie County, New York. *In* New York State Geol. Assn. Guidebook, 46th Ann. Mtg., SUNY College at Fredonia, p. G1-G15.

_________ and Baird, G. C. 1982. Upper Moscow-Genesee stratigraphic relationships in western New York: evidence for regional erosive beveling in Late Middle Devonian. *In* New York State Geol. Assn. Guidebook, 54th Ann. Mtg., SUNY at Buffalo, p. 19-63.

_________, Dick, V.B., and Baird, G.C. (in press). Taphonomy and paleoecology of Middle Devonian (Hamilton Group) dark gray and black shales from western New York. *In* Kauffman, E.G., ed., Paleontology of Organic-Rich Strata. Hutchinson and Ross.

_________, Baird, G.C., Gray, L.M., and Savarese, M.L. 1983. Middle Devonian (Givetian) coral associations of western and central New York State. *In* Sorauf, J.E. and Oliver, W.A., Jr., eds., Silurian and Devonian Corals and Stromatoporoids of New York. Int. Assn. for the Study of Fossil Cnidaria, 4th Int. Symp., Washington, D.C., p. 65107.

_________, Speyer, S.E., and Baird, G.C. (this volume). Storm generated sedimentary units: tempestite proximality and event, stratification in the Middle Hamilton Group of New York.

Brown, P.R. 1966. Pyritization in some molluscan shells. J. Sediment. Petrol. 36:149-1152.

Byers, C.W. 1974. Shale fissility: relation to bioturbation. Sedimentology 21:479-484.

_________ 1977. Biofacies patterns in euxinic basins: a general model. *In* Enos, P. and Cook, H.E., eds., Deep Water Carbonate Environments. Soc. Econ. Paleont. and Mineral. Spec. Publ. 25:5-17.

Caster, K.E. 1934. The stratigraphy and paleontology of northwestern Pennsylvania. Bull. Amer. Paleont. 21, 185 p.

Clarke, J.M. 1885. On the higher Devonian faunas of Ontario County, New York. U. S. Geol. Surv. Bull. 16:1-28.

Clarke, G.R. and Lutz, R.A. 1980. Pyritization in the shells of living bivalves. Geology 8:268-271.

Cleland, H.F. 1903. A study of the fauna of the Hamilton Formation of the Cayuga Lake section in central New York. U.S. Geol. Surv. Bull. 206, 111 p.

Cloud, P.E. 1948. Assemblages of diminutive brachiopods and their paleobiological significance. J. Sediment. Petrol. 18:56-60.

Cluff, R.M. 1980. Paleoenvironment of the New Albany Shale Group (Devonian-Mississippian) of Illinois. J. Sediment. Petrol. 50:767-780.

Cooper, G.A. 1930. Stratigraphy of the Hamilton Group of New York, Parts 1 and 2. Amer. J. Sci. 19:116-134.

_________ 1957. Paleoecology of the Middle Devonian of eastern and central United States. Geol. Soc. Amer. Mem. 67:249-278.

Curtis, C.D. and Spears, D.A. 1968. The formation of sedimentary iron minerals. Econ. Geol. 63:57-270.

Dennison, J.M. and Head, J.W. 1975. Sea level variations interpreted from the Appalachian Basin Silurian and Devonian. Amer. J. Sci. 275:1089-1120.

Dick, V. 1982. Taphonomy and depositional environments of Middle Devonian (Hamilton Group) pyritic fossil beds in western New York. Master's thesis, University of Rochester, 115 p.

Dickson, J.A.D. 1966. Carbonate identification and genesis as revealed by staining. J. Sediment. Petrol. 36:491-505.

Domagala, M.A. and Selznick, M.R. 1979. Paleontology and stratigraphy of the Ledyard Shale (Middle Devonian) at Spring Creek, Alden, New York. Rochester Acad. Sci. Spec. Pub. #1, 25 p.

Einsele, G. and Mosebach, R. 1955. Zur petrographie, Fossilerhaltung und Entstehung der Gestein des Posidonienschiefers in Schwabischen Jura. Neues Jb. Geol. Palaontol. Abh. 101:319-340.

Ettensohn, F.R. and Barron, L.S. 1981. Depositional model for the Devonian-Mississippian black shales of North America: a paleoclimatic-paleogeographic approach. *In* Roberts, T.G., ed., Geol. Soc. Amer. Guidebook, 1981 Field Trip, Cincinnati, Amer. Geol. Inst. 2:344-361.

Fisher, D.W. 1951. Marcasite fauna in the Ludlowville Formation of western New York. J. Paleont. 15:61-72.

Fisher, D.C. 1977. Mechanism and significance of enrollment in xiphosurans (Chelicerata, Merostomata). Geol. Soc. Amer. Abstr. with Programs 9:264-265.

Frey, R.W. 1970. Trace fossils of Fort Hays Limestone Member of Niobrara Chalk (Upper Cretaceous), west-central Kansas. Univ. Kansas Paleo. Contrib. Article 53, 41 p., 10 pls.

Goldhaber, M.B. and Kaplan, I.R. 1974. The sulfur cycle. *In* The Sea, v. 5, (Marine Chemistry) p. 569-655.

Grasso, T.X. 1970. Paleontology, stratigraphy and paleoecology of the Ludlowville and Moscow Formations (upper Hamilton Group) in central New York. *In* New York State Geol. Assn. Guidebook, 42nd Ann. Mtg., SUNY College at Cortland, p. D1D18.

Gregoire, C. 1964. On submicroscopic structure of the *Nautilus* shell. Inst. Royal des Sci. Nat. de Belgique Bull. 38:1-71.

Hallam, A. 1965. Environmental causes of stunting in living and fossil marine benthonic invertebrates. Paleontology 8:132-155.

Howarth, R.W. 1979. Pyrite: its rapid formation in a salt marsh and its importance in ecosystem metabolism. Science 203:49-51.

Hudson, J.D. 1978. Pyrite in ammonite shells and in shales. Neues Jb. Geol. Paleont. 157:190-193.

_________ 1982. Pyrite in ammonite-bearing shales from the Jurassic of England and Germany. Sedimentology 27(2):639-667.

_________ and Palframan, D.F.B. 1968. The ecology and preservation of the Oxford Clay fauna at Woodham, Buckinghamshire. Quart. J. Geol. Soc. London 124:378-418.

Izard, J.E. and Clemency, C.V. 1967. X-ray study of the sedimentary pyrite of western New York. J. Sediment. Petrol. 37:221-225.

Kalliokowski, J. 1966. Diagenetic pyritization in three sedimentary rocks. Econ. Geol. 61:872-885.

Kammer, T.W. 1981. Prodeltaic communities associated with submarine fan valleys, Borden Formation (Mississippian), Kentucky and Indiana. Geol. Soc. Amer. Abstr. with Programs 13:1-482.

Kaplan, I.R., Emery, K.O., and Rittenberg, S.C. 1963. The distribution and isotopic abundance of sulfur in recent marine sediments of southern California. Geochim. et Cosmochim. Acta 27:297-332.

Kepferle, R.C. 1977. Stratigraphic, petrology, and depositional environment of the Kenwood Siltstone Member, Borden Formation (Mississippian), Kentucky and Indiana. U.S. Geol. Surv. Prof. Paper 1007:1-49.

Krauskopf, K.B. 1967. Introduction to Geochemistry. McGraw Hill, New York, 720 p.

Loomis, F.B. 1903. The dwarf fauna of the pyrite layer at the horizon of the Tully Limestone in western New York. New York State Mus. Bull. 69:892-920.

McAlester, A.L. 1968. Type species of Paleozoic nuculoid bivalve genera. Geol. Soc. Amer. Mem. 105, 143 p.

McCollum, L. 1980. Paleoecology of the Ledyard Shale Member (Middle Devonian: Givetian) of New York. Doctoral dissertation, SUNY at Binghamton, 143 p.

Morris, K.A. 1980. Comparison of major sequences of organic-rich mud deposition in the British Jurassic. J. Geol. Soc. London 137:57-170.

Muller, A.H. 1979. Fossilization (Taphonomy). *In* Robison, R.A. and Teichert, C., eds., Treatise on Invertebrate Paleontology. Geol. Soc. of Amer. and Univ. Kansas Press, Part A, A2-A78.

Nussman, D.G. 1975. Paleoecology and pyritization. Univ. of Michigan Mus. of Paleontol., Papers on Paleontol. 8:173-223.

Park, W.C. and Weiss, A.J. 1972. Diagenetic sulfides in fossiliferous Leicester Marcasite (upper Hamilton Group) from Devonian of New York. Geol. Soc. Amer. Abstr. with Prog. 4:1-38.

Potter, P.E., Maynard, J.B., and Pryor, W.A. 1980. Sedimentology of Shale. Springer-Verlag, Inc., New York, 306 p.

Reaves, C.M. 1984. The migration of iron and sulfur during the diagenesis of marine sediments. Doctoral dissertation, Yale Univ., 413 p.

Reyment, R.A. 1971. Introduction to Quantitative Paleoecology. Elsevier Publ. Co., New York, 226 p.

Rhoads, D.C. 1970. Mass properties, stability and ecology of marine muds related to burrowing activity. *In* Crimes, T.P. and Harper, J.C., eds., Trace Fossils, Seel House Press, Liverpool, p. 391-406.

__________ 1974. Organism-sediment relations on the muddy seafloor. *In* Barnes, H., ed., Ann. Rev. Mar. Biol. 12:263-300. Allen & Unwin.

__________ and Morse, J.W. 1971. Evolution and ecological significance of oxygen deficient marine basins. Lethaia 3(4):3-428.

Richards, F.A. 1965. Anoxic basins and fjords. *In* Riley, J.P. and Skirrow, G., eds., Chemical Oceanography, Academic Press, New York, p. 611-645.

Rickard, D.T. 1975. Kinetics and mechanisms of pyrite formation at low temperatures. Amer. J. Sci. 275:635-652.

Rickard, L.V. 1981. The Devonian system of New York State. *In* Oliver, W.A., Jr. and Klapper, G., eds., Devonian Biostratigraphy of New York. Subcom. on Dev. Strat., Intl. Union of Geol. Sci., Washington, D.C., p. 5-22.

Rudwick, M.J.S. 1970. Living and fossil brachiopods. Hutchinson Univ. Press, London, 199 p.

Savrda, C.E., Bottjer, D.J., and Gorsline, D.S. 1984. Development of a comprehensive euxinic biofacies model: evidence from Santa Monica, San Pedro and Santa Barbara Basins, California continental borderland. Amer. Assn. Petroleum Geol. Bull. 67(3):544.

Stanley, S.M. 1970. Relation of shell form to life habits in the Bivalvia (Mollusca). Geol. Soc. Amer. Mem. 125, 126 p.

Stein, W.E. 1981. Reinvestigation of *Arachnoxylon kopfii* from the Middle Devonian of New York State, USA. Palaontographica, Bd. 177, Abt. B, p. 90-117, 14 pls.

Stenzel, H.B. 1964. Living Nautilus. *In* Moore, R.C., ed., Treatise on Invertebrate Paleontology. Geol. Soc. of Amer. and Univ. Kansas Press, Part K, K59-K93.

Sutton, R.G., Bowen, Z.P. and McAlester, A.L. 1970. Marine shelf environments of the Upper Devonian Sonyea Group of New York. Geol. Soc. Amer. Bull. 81:2975-2992.

Thayer, C.W. 1975. Morphologic adaptations of benthic invertebrates to soft substrata. J. Mar. Res. 33:177-189.

Thompson, J.B., Mullins, H.T., Newton, C.R., and Vercoutere, T.L. 1985. Alternative biofacies model for dysaerobic communities. Lethaia 18:167-179.

Thomsen, E. and Vorren, T.O. 1984. Pyritization of tubes and burrows from Late Pleistocene continental shelf sediments off north Norway. Sedimentology 31:481-492.

Towe, K.M. and Grimm, R.E. 1963. Variations in clay mineralogy across facies boundaries in the Middle Devonian (Ludlowville) of New York. Amer. J. Sci. 261:839-861.

Weller, J.M. 1959. Compaction of sediments. Amer. Assn. Petroleum Geol. Bull. 43:273-310.

Williams, H.S. 1903. The correlation of geologic faunas. U. S. Geol. Surv. Bull. 210.

Wygant, G.T. this volume. Deposition and early diagenesis of a middle marine shale: Ludlowville Formation, western New York.

Zangerl, R. 1971. On the geologic significance of perfectly preserved fossils. Proc. North Amer. Paleo. Conv. V, p. 1207-1222.

__________, Woodland, B.G., Richardson, E.S., and Zachry, D.L. 1969. Early diagenetic phenomena in the Fayetteville Black Shale (Mississippian) at Arkansas. Sediment. Geol. 3:87-120.

APPENDIX

Locality Register

1) Angola and Eden Quadrangles — Lake Erie shoreline bluffs north and south of Eighteenmile Creek and Pike Creek.
2) Eden Quadrangle — shale banks along Eighteenmile Creek on Rt. 5 approximately 2 km west of Lake View Road.
3) Buffalo Southeast Quadrangle — small creek flowing behind the WKBW radio towers near the intersection of Robin Lane and Big Tree Road in Hamburg, New York.
4) Buffalo Southeast Quadrangle — abandoned shale borrow pit off Big Tree Road near the intersection of Big Tree Road and Bay View Road and west of Rush Creek.
5) Buffalo Southeast Quadrangle—stream banks of Smoke Creek, 2. 9 km southwest of Ebenezer, New York (north branch of creek) and of Abbott Road in Windom, New York (south branch).
6) Orchard Park Quadrangle — shale banks of Cazenovia Creek from bridge at Northrup Road to 2 km upstream. Locality is 0. 8 km west of Springbrook, New York.
7) East Aurora Quadrangle — shale banks of Buffalo Creek south of Bullis Road bridge in the town of Elma.
8) East Aurora Quadrangle — banks of Little Buffalo Creek northwest of intersection of Town Line Road and Bullis Road in the Town of Elma.
9) East Aurora Quadrangle — banks of tributary to Cayuga Creek, off Cayuga Creek Road southwest of Alden, New York.
10) Corfu Quadrangle — shale banks of Spring Creek 0. 3 – 0. 5 km northeast of Alden, New York.
11) Corfu Quadrangle — shale banks in Elevenmile Creek north (downstream) of US Rt. 20 overpass, 2. 4 km west of Darien, New York.
12) Alexander Quadrangle — banks of Bowen Creek near Broadway Road, 0. 3 km west of the Town of Alexander.
13) Batavia South Quadrangle — banks of Little Tonawanda Creek, south of U. S. Rt. 20 bridge over the creek.
14) Batavia South Quadrangle — spoil along banks of Delaware, Lackawana and western railroad tracks near Francis Road overpass, 2. 5 km east of East Alexander, New York.
15) Batavia South Quadrangle — east-facing shale pit bank on Bethany Center Road, 3.0 km north of Bethany Center, New York.
16) Stafford Quadrangle — banks of White Creek between US Rt. 20 and Rt. 63, 2. 0 km west of Texaco Town.

Storm-Generated Sedimentary Units: Tempestite Proximality and Event Stratification in the Middle Devonian Hamilton Group of New York

Carlton E. Brett and Stephen E. Speyer
Department of Geological Sciences
The University of Rochester

Gordon C. Baird
Department of Geology
State University College at Fredonia

ABSTRACT

Recent advances in the understanding of storm processes permit reevaluation of numerous ancient sedimentary units as the direct or indirect effects of storm wave impingement on the sea floor. Beds reflecting storm-generated deposition, tempestites, occur at many levels of the Middle Devonian Hamilton Group of western and central New York.

Thick, tabular calcarenites, such as the Tichenor-Portland Point limestones, record multiple storm events, obscured by successive reworking of bioclasts in shallow (10-20 m) marine settings. Such layers yield evidence for both long term condensation and sudden pulses of burial. These calcarenites appear to grade basinward into sequences of discrete storm layers. The latter, in turn, display a bathymetrically-related "proximality spectrum" ranging from sandy/shelly coquinites, through calcisiltites/siltstones (medial tempestites) to more distal mud burial layers. Most Hamilton shell beds are lenticular, parautochthonous coquinites; some exhibit evidence for more than one episode of storm-wave reworking during an interval of low background sedimentation. Siltstones and calcisiltites, displaying grading and fine planar to hummocky lamination, probably were deposited from waning density currents charged with storm-winnowed silts. The rarity of these beds in the Hamilton, as compared with many older stratigraphic units is probably related to pervasive bioturbation by deeper infauna (e.g., *Zoophycos*) not present during the Early Paleozoic.

Distal tempestites are represented by "smothered bottom" fossil assemblages (e.g., "trilobite" beds). Laterally persistent carbonate lenses and concretionary beds apparently represent diagenetic enhancement of such buried horizons. These beds frequently display taphonomic features indicative of rapid burial (e.g., complete trilobites and pelmatozoans, bivalves in burrow position). Furthermore, fossils within concretionary carbonates show less severe compactional deformation and biogenic disturbance than those within synjacent shales. These observations imply early diagenetic carbonate cementation of muds. We suggest that anaerobic decay of suddenly-buried benthos locally raised the pH of interstitial waters, thereby favoring selective precipitation of calcium carbonate. The proximality model relates various types of storm deposits to relative depth: winnowed shell beds are mainly restricted to a zone of direct scouring of the sea floor by storm waves, whereas, calcisiltites and mud layers represent deposition from suspension clouds downslope from storm wave base. The nearly complete "proximality spectrum" can be observed along single storm beds associated with the northern margin of the Staghorn coral bed (Ludlowville Formation) at Skaneateles Lake.

This example confirms the utility of storm beds in inferring relative depth. Moreover, because of their single event origin, tempestites are isochronous and, therefore, represent important stratigraphic markers.

INTRODUCTION

A major contribution to the understanding of shallow epeiric sea facies is the recent documentation of the importance of episodic or catastrophic events, rare in terms of human lifetimes, but inevitable and common in geologic time (Ager, 1974; Dott, 1983). Sedimentary packages resulting from such processes include both a)single-event beds, (e.g., ash fall layers, turbidites and storm deposits) representing essentially isochronous units deposited during time spans ranging from hours to a few days, and b)more complex multiple-event beds that may represent the mixing of several single-event deposits, or be manifest as omission surfaces or condensed beds. Such episodic sedimentary units are distinct from background sequences that reflect more nearly continuous, generally slow, sediment accumulation.

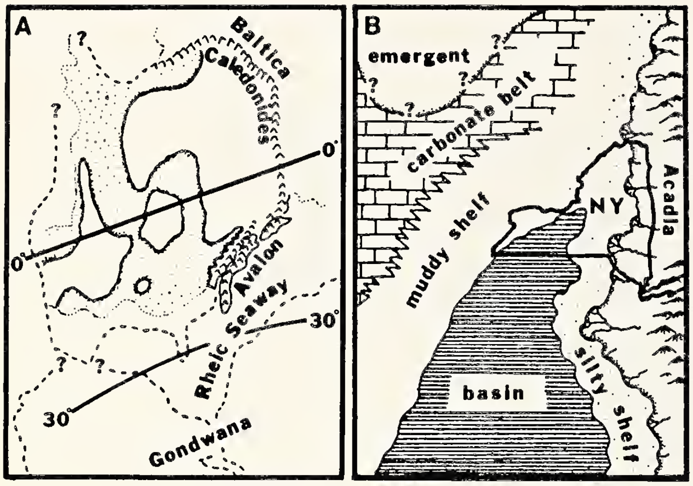


Figure 1 Depositional setting of the Hamilton Group. A) Paleogeography of North America during Middle Devonian (Givetian) time; stippled area represents epeiric seaway; note Acadian belt between Avalon and North America. B) Northeastern North America region showing northern end of Appalachian Basin extending into area of New York State, bordered on the southeastern side by deltaic complex and on the north-northwest by muddy carbonate shelf regions.

Recently, a number of criteria for recognition of episodic storm-generated event deposits (tempestites) have been developed from both Recent shelf sediments (Reineck and Singh, 1972; Ball and others, 1967) and their ancient counterparts (Goldring and Bridges, 1973; Kumar and Sanders, 1976; Cant, 1980; Kreisa, 1981; Kreisa and Bambach, 1982; Aigner, 1979, 1982). The application of a "tempestite model" to ancient depositional environments requires detailed reevaluation of sedimentary sequences on a bed-by-bed basis. The approach promises new insights into shelf depositional processes, in much to same way that the recognition of turbidites (Bouma, 1962) revolutionized the study of slope-and-rise environments.

The concept of proximality, originally used with turbidites (cf.Walker, 1967) has been recently extended to the interpretation of tempestite sequences. Because the intensity of storm effects decreases away from shore, proximal storm deposits are characteristically coarser and thicker than distal tempestites (Aigner, 1982, p. 187; Aigner and Reineck, 1982). Repeatedly reworked lag deposits along coastal shorefaces (cf.Kumar and Sanders, 1976) represent amalgamated storm layers; these diverge basinward into discrete storm layers, which represent single depositional events. Tempestite lithofacies (sensu Kelling, in Ager, 1973) are characterized by thin (3-10 cm), often graded, beds which alternate with background sediments; the most distal storm layers have been termed mud tempestites (Aigner and Reineck, 1982). Although a substantial literature documents the character of more proximal storm deposits (i.e., shallow shelf), event deposition in deeper, offshore settings remains poorly understood. We believe that many previously reported "smothered bottom" fossil assemblages (e.g., Hallam, 1961; Nussman, in Kesling and Chilman, 1975; Brett and Eckert, 1982; Speyer and Brett, 1982) are the downslope expression of tempestite facies described by Aigner (1979, 1982, 1985), Kreisa (1981), and others.

Kreisa (1981) and Duke (1982) have suggested a direct correlation between the incidence of probable storm-generated features (e.g., hummocky cross-stratification) and paleolatitude. Duke (1982) notes that contemporary climate belts resemble those which prevailed in the Paleozoic, although Marsaglia and DeVries-Klein (1983) caution that storm stratigraphy and paleogeography may not always be closely related.

Recently proposed paleogeographic reconstructions (e.g., Oliver, 1976; Irving, 1980) place the northern Appalachian Basin during Middle Devonian time at about 10° south latitude (Fig. 1), well within the region of probable hurricanes or tropical storm occurrence. Hamilton Group lithofacies along the New York State outcrop belt should record the effects of episodic storm deposition (Fig. 1), first because of the above paleogeographic factor, and second because of the shallow character of many Hamilton facies. Moreover, depositional features of the contemporaneous Mahantango Formation, and slightly older (Emsian-Eifelian) Esopus and Onondaga formations in Pennsylvania and New York have previously been attributed to storm processes (Goldring and Bridges, 1973; Duke, 1982).

Our observations indicate that a number of beds in the Hamilton Group of west-central New York also represent storm deposits; furthermore, specific depositional characteristics are apparently facies related. These include simple and complex shell beds or coquinites (Rich, 1951), particularly those showing associated well-preserved fossils, and thin, finely laminated siltstones and calcisiltites. Moreover, numerous thin, regionally persistent horizons of extremely well-preserved fossils (*Konservat-lagerstatten, sensu* Seilacher, 1970) occur within the Hamilton sequence; some of these have been recorded in the literature as "crinoid beds" (Goldring, 1923); pyrite nodule horizons (Fisher, 1951), and "trilobite beds" (Grabau, 1898-1899). We interpret such units as smothered bottom assemblages indicative of rapid blanketing of extensive sea floor areas by storm-generated mud clouds (Brett and Baird, 1984; Speyer and Brett, 1984). Significantly, many such beds display the effects of a selective early diagenetic overprint which facilitates their recognition and mapping within thick, uniform "background" deposits.

In the present paper we describe several Hamilton layers, representing both single and multiple storm events, and present an interpretation of the genesis and bathymetric significance of each. A study of a single outcrop exposing a variety of storm beds is presented to further illustrate the storm bed proximality model. Many of the layers, here interpreted as tempestites, are widely traceable and provide significant regional markers. Hence, they are also important as potentially isochro-

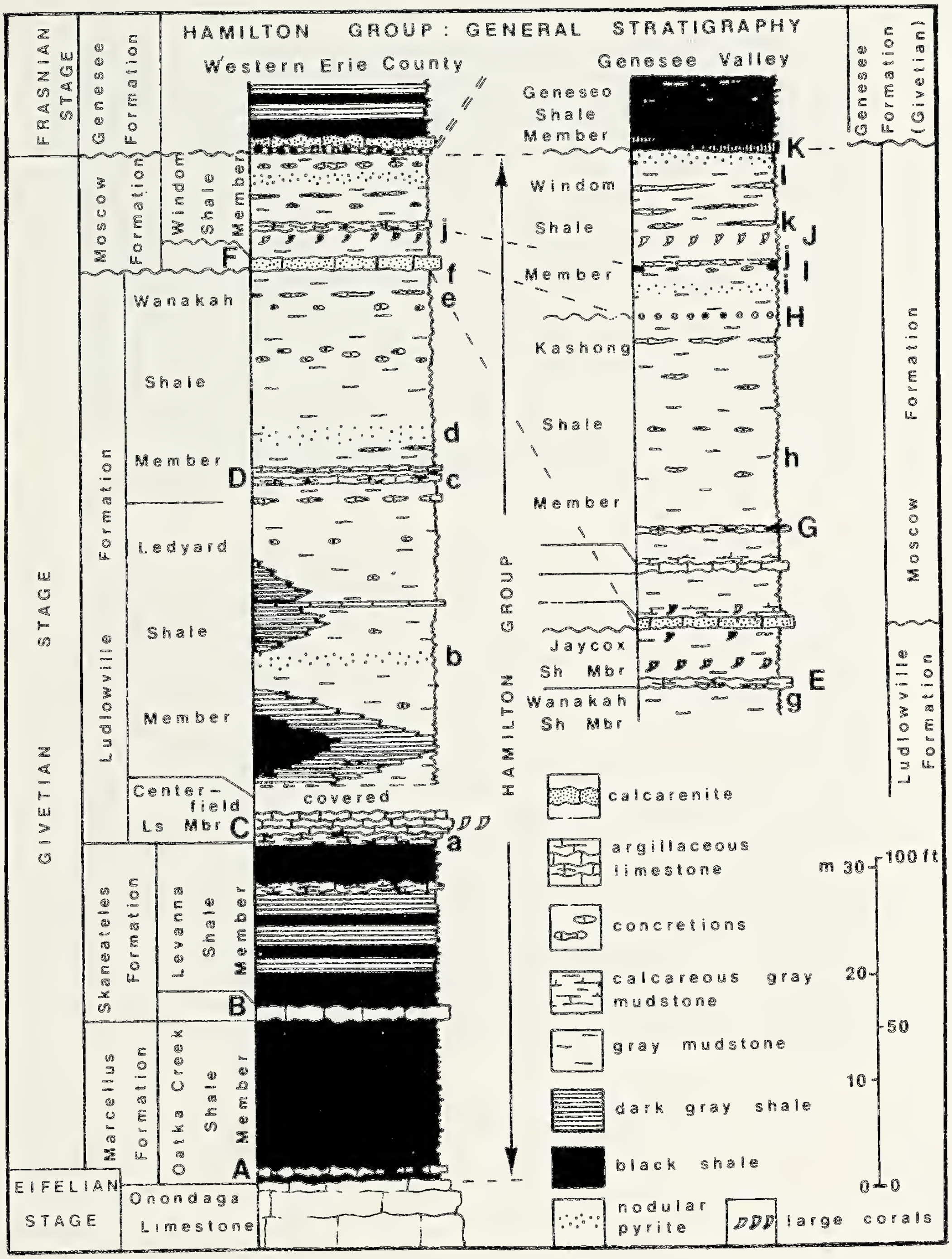


Figure 2 Generalized stratigraphic column for the Hamilton Group in western New York. Inset shows details of stratigraphy in the Genesee Valley region; note additional units absent at Erie County 90 km to the west. Positions of major event beds discussed in text are designated by letters; multiple event, condensed, beds (capital letters) include: A) Cherry Valley Limestone; B) Stafford Limestone; C)Centerfield Limestone; D) "*Pleurodictyum* beds"; E) Hills Gulch bed; F) Tichenor Limestone; G) Menteth Limestone; H) upper Kashong phosphatic pebble bed; I) Bay View Coral bed; J) Fall Brook coral beds and K) Leicester Pyrite bed.
Event beds are as follows (by member): Centerfield Member: a) Browns Creek (trilobite) bed. Ledyard Shale Member: b) Alden Pyrite bed. Wanakah Shale Member: c) Murder Creek (trilobite) bed; d) Francis Road pyrite beds; e) Spring Brook concretionary beds; f) "*Stictopora*" beds (crinoid layers). Jaycox Shale Member: g) basal bivalve bed. Kashong Shale Member: h) "*Retsof*" crinoid-bryozoan lenses. Windom Shale Member: i) Little Beards Creek pyrite bed; j) Smoke Creek (trilobite) bed; k) middle Taunton beds; 1) upper Taunton (trilobite) beds.

nous key beds. Furthermore, the proximality spectrum of storm beds within the Hamilton Group provides a useful parameter for inferring relative depth in otherwise indistinguishable background mudstones.

GEOLOGIC SETTING

The Hamilton Group (Middle Devonian: Givetian) is a predominantly detrital sediment package which records terrigenous sediment influx from major uplifted regions in New England and in the mid-Atlantic region. This sequence, divisible into four formations, is composed predominantly of shale and mudstone deposits with minor beds and intervals of limestone or sandstone (Fig. 2); it records shelf and basin margin marine settings near the northern termination of an inland sea (Fig. 1). The northern and western boundaries of this sea bordered low-relief cratonic regions which supplied relatively little detritus to the basin as compared to the uplifted terranes; this accounts for thin Hamilton deposits in western New York (Dennison and Head, 1975) and the eastward thickening clastic wedges of equivalent sediments in central and eastern New York (Cooper, 1957; Rickard, 1981).

Hamilton deposits in the study area of western and central New York record subtidal marine shelf to basin environments along the northern margin of the Appalachian Basin (see Baird and Brett, 1981, 1983; Brett and Baird, 1982). Although strata exposed in western New York counties display remarkable continuity in stratigraphic spacing which suggests stable platform conditions, outcrop data in the Finger Lakes region and from the subsurface of western New York indicate a southward transition to basinal facies at most levels (Baird and Brett, 1981, 1983). This is reflected by a southward gray to black shale facies change, southward transition of high energy carbonate deposits (e.g., Centerfield, Tichenor limestones) to shaley micrites, and northward appearance and expansion of stratigraphic discontinuities. Basinal facies intercept the outcrop belt in the Finger Lakes Region (see Savarese and others, this volume); this northeast-trending trough provides an important view of deeper water environments and allows regional examination of storm beds in the context of a depth model. It is evident that widespread water depth fluctuations have been superimposed on Hamilton paleogeographic facies patterns; at least twenty regressive-transgressive cycles of varying magnitude and completeness are represented within the four Hamilton formations (Fig. 2). As predicted by the proximality model, these would have had an important effect on the character of storm-generated deposits.

The rocks discussed herein include marine sediments deposited in shelf settings ranging from high energy subtidal regimes to anaerobic basin conditions. The great bulk of the Hamilton sequence is composed of shales and mudstones, recording platform and gentle slope deposition below normal wave base. Hence, the record of major storms is often preserved where the event signature has not been erased by the activity of burrowing organisms. Units deposited at or near wave base are infrequent but include thin limestones or sandstones with characteristic hummocky cross-stratification (cf. Harno and others, 1975; Dott and Bourgeois, 1982). Siltstone and calcisiltite intervals are characteristically heavily bioturbated and typically appear uniform and often massive in outcrop. Open shelf mudstones often appear thick and uniform as well, but careful examination reveals the presence of widespread marker beds, some occurring as single bedding planes. Such "event" horizons often record storm-induced bottom smothering by suspended clouds of mud and fine silt; these beds, representing marginal aerobic and dysaerobic Hamilton environments, are a central topic of the present paper.

DISTRIBUTION AND REGIONAL CHARACTER OF HAMILTON STORM BEDS

Hamilton Shell Beds

Portions of the Hamilton sequence, particularly in the Ludlowville and Moscow formations, contain abundant thin (1.5 cm thick) shell beds or coquinites (see Boardman, 1960, for discussion). Although they are most common within moderately fossiliferous, bioturbated mudstone and in dark-gray to black fissile shales. Most Hamilton shell horizons are thin and lenticular and may be termed simple shell beds (cf.Kidwell and Jablonski, 1983). A smaller number of Hamilton shell-rich beds appear tabular, and are more widely mappable; these thicker shell beds reflect more complex depositional histories and represent longer intervals of time than do simple shell beds. Although simple and complex shell beds probably represent end members of a continuum, these types of shell beds will, for convenience, be considered separately in the following sections. It should be emphasized that even thin, seemingly simple shelly horizons may have regionally widespread distribution, complex depositional histories, and spatially variable appearance between outcrops.

Complex Shell Beds; Amalgamated Storm Layers

Certain Hamilton calcarenite and sandy encrinite units, generally occurring at or near the regressive centers of sedimentary cycles (e.g., axial portions of the Mottville, Centerfield and Portland Point members and their clastic equivalents), exhibit features which indicate multiple reworking and persistent winnowing of sediments. Such complex shell-sand layers are termed amalgamated beds (*sensu* Goldring and Aigner, 1982; Aigner and Reineck, 1982). By nature they are regionally widespread and traceable for distances of tens to hundreds of kilometers along the outcrop belt.

The Tichenor Limestone and coeval basal Portland Point Member in western and central New York State (Fig. 2; see Baird, 1979) offers an excellent example of this type of deposit. These units are normally represented by single, tabular beds, 0.5 to 1.5 m thick, composed chiefly of disarticulated, com-

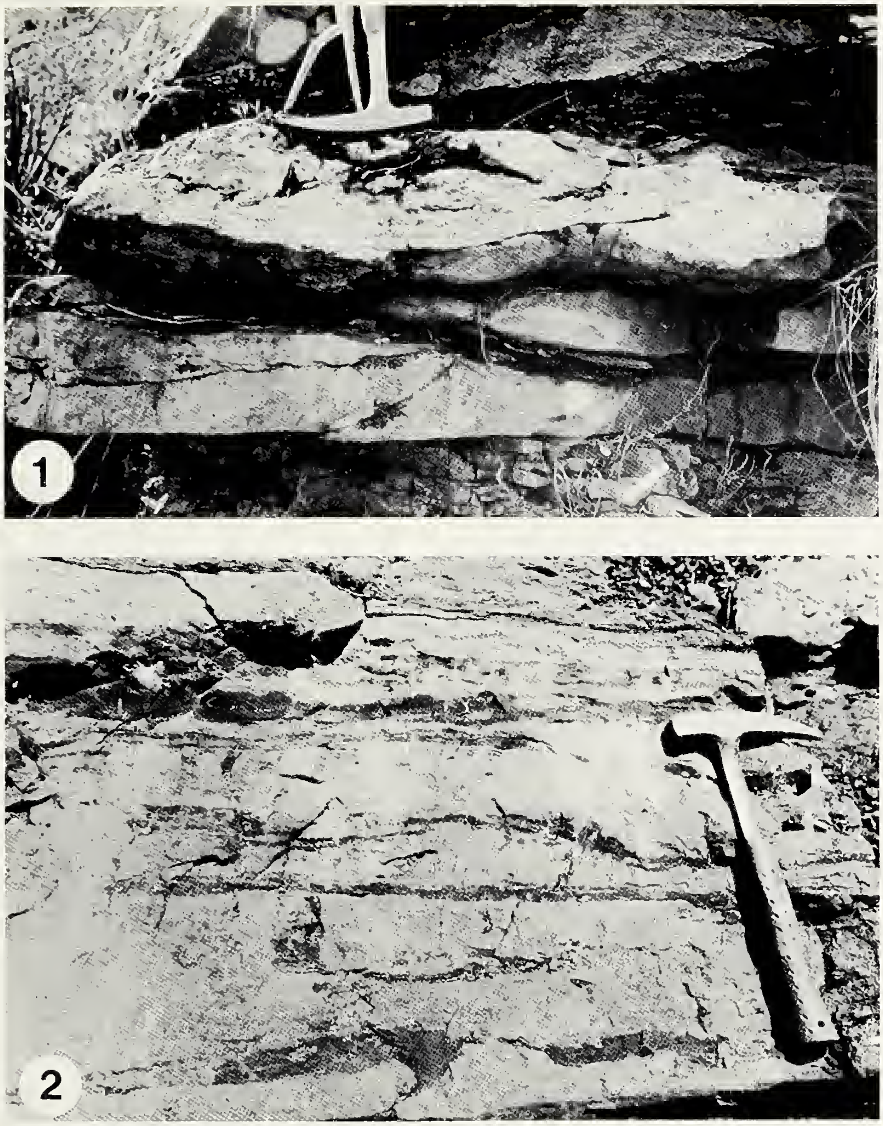


Figure 3 1) Outcrop of Tichenor Limestone showing scour-and-fill channel. Sheldrake Creek, west of Sheldrake, Cayuga County, New York. 2) Portland Point limestone; note lenticular bedding with hummocky ripples; troughs are filled with mud; Portland Point, Cayuga Lake east shore, Lansing, Tompkins County, New York.

minuted crinoidal debris and containing rugose and tabulate corals and large spiriferid brachiopods. The Tichenor and similar Hamilton limestones clearly are condensed deposits which grade laterally into thick mudrock sequences (Baird, 1979; Baird and Brett, 1981). Conodont biostratigraphy further indicates that the Tichenor Limestone in western New York records a long time span, including parts of two conodont subzones (see Klapper, 1981). Furthermore, such condensed units display internal features diagnostic of long term reworking and minimal deposition, including: a)abrasion, bioencrustation and biocorrosion of shells, and b)reworking of early diagenetic structures (e.g., phosphate nodules) and prefossilized material. Nonetheless, these beds also contain evidence for episodes of reworking involving shifting of sediments and rapid burial. Thus, for example, the Tichenor locally contains well-preserved remains of fragile skeletons together with highly corroded, roll-faceted fossils. Complete crinoids, blastoids and trilobites occur commingled with abraded, roll-fragments of the brachiopod *Spinocyrtia* (Fig. 4). These last, obviously are residual, reworked bioclasts, whereas fossils with multielement skeletons such as crinoids and trilobites must be buried permanently within a few hours following death if they are to be preserved intact (Meyer, 1971; Liddell, 1975; Plotnick, 1984).

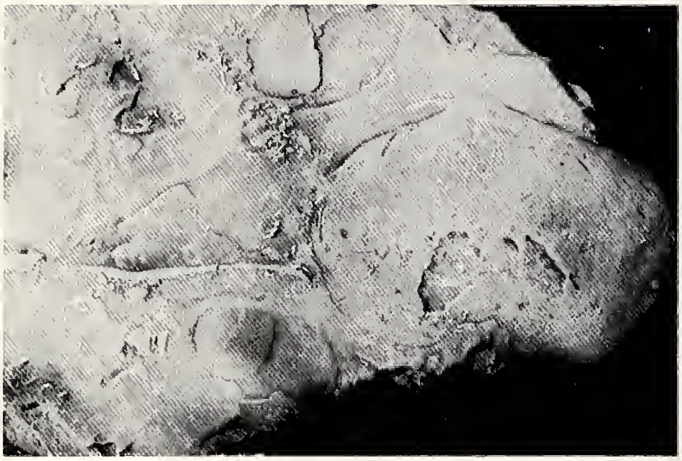


Figure 4 Heavily corroded and breached pedicle valve of *Spinocyrtia*; note uncorroded valves of *Centronella*, bryozoa and other fossils, in direct association with *Spinocyrtia*; "*Rhipidomella-Centronella* bed," Kashong Shale Member; Kashong Glen, near Bellona, Yates County, New York, X 1.5.

Sedimentary structures in the Tichenor also record the imprint of episodic reworking by storm processes. The base of the limestone is everywhere sharply erosional on the underlying beds. The Tichenor demonstrably oversteps older units both east and west of the Seneca Lake region, indicating that this contact records long-term submarine erosion (Baird, 1979; Baird and Brett, 1981). Nonetheless the basal contact also locally displays evidence for episodic rapid scouring and shifting of sediment.

At many localities, elongate, prodlike structures protrude downward as much as 10 cm into subjacent shales (Fig. 5). In some examples these structures represent casts of large burrowing traces. Some traces show apparent scratch marks which suggest an arthropodan origin; "megaburrows" of nearly identical dimensions also have been observed at similar discontinuity surfaces in the Middle Devonian Hungry Hollow Limestone in Ontario (Landing and Brett, in prep.), and the Tully Limestone in New York (Heckel, 1973). Preservation of delicate scratch marks on sole castings requires that the underlying muds were firm and that the burrows were filled by carbonate sediment shortly after their production. The fact that they are often filled with relict sediments, including phosphatic granules and pebbles and abraded fragments of fish bone, suggests episodic tractional movement of remanié debris blankets, perhaps as low bars which migrated across a regionally scoured substrate. Evidently, the last burial episode in a particular region was sufficiently rapid to preserve delicate traces intact; this sediment cover presumably shielded the scour surface from future disturbance.

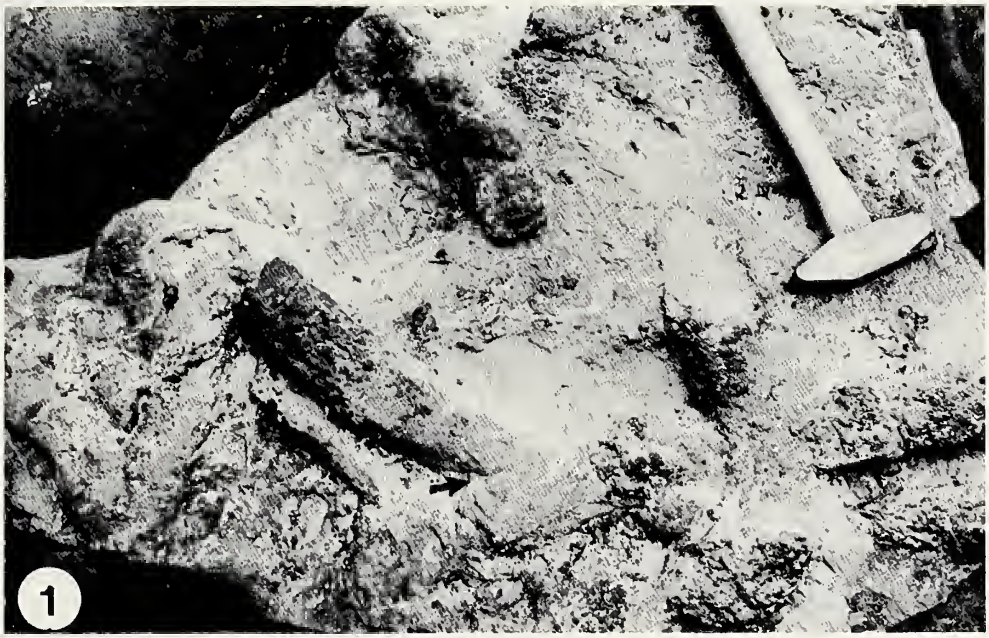
In other areas the underface prods of the Tichenor are irregularly rounded and lobate, closely resembling gutter casts described by Aigner and Futterer (1978). These prods display shapes and sizes quite similar to those noted for the megaburrows described above, suggesting that the biogenic furrows served as loci to concentrate surface scour by turbulence associated with the transport of burial sediment. Gutter casts have been associated with proximal storm sand deposition in both ancient and recent shallow sea floor sequences (Aigner and Reineck, 1982). Previous studies (Aigner and Futterer, 1978) have associated such structures with storm deposition in shallow-water areas, generally 20 m or less in depth.

That the condensed encrinite and sandy sediments were reworked after storm deposition is indicated by cross cutting of internal sedimentary structures. The Tichenor Limestone, locally (as at Sheldrake Creek, Fig. 3) displays large (up to 0.5 m deep), channel-like scours which truncate older stratification; these channels may be filled with cross laminated debris layers separated by thin gray shale partings. Such scour-and-fill features provide excellent evidence for multiple episodes of erosion and deposition within condensed beds (cf.Goldring and Aigner, 1982).

Simple Shell Beds: Storm-Winnowed Coquinites

Thin lenticular shell beds are particularly characteristic of medium-gray mudstones containing variably diverse faunas. Such beds have been observed in association with lower diversity ambocoeliid biofacies (see Brett, Dick and Baird, in press, for discussion), but occur more commonly with higher diversity brachiopod-, coral- and bryozoan-bearing facies such as the *Stereolasma*, *Pleurodictyum*, and *Heliophyllum* coral associations (see Brett and others, 1983, for more detailed descriptions). Shell beds, interpretable as tempestites, have been reported from numerous fossil-rich zones including the *"Pleurodictyum"* and *"Demissa"* beds of the Ludlowville Wanakah Shale (Boardman, 1960; Miller, 1982), portions of the Kashong Shale and the Bay View and Taunton beds of the Moscow Windom Shale (Baird and Brett, 1983). Shell beds of this type also occur within similar lithofacies of the lower Hamilton Group exposed in central New York.

Simple shell beds are typically thin, ranging from a single layer of enriched fossil debris, only a few millimeters in thickness, to lenses 1-10 cm thick; they are most commonly lenticular, pinching out within a few meters along a given outcrop. In some examples "pods" of skeletal debris several centimeters thick pinch down laterally into thin, shelly stringers which yield only scattered specimens of fossils commonly represented in the thicker lentils. Several shell beds (e.g., those from the Kashong Shale) are lenticular, displaying periodic fluctuations in thickness; these resemble starved ripples (cf. Potter,



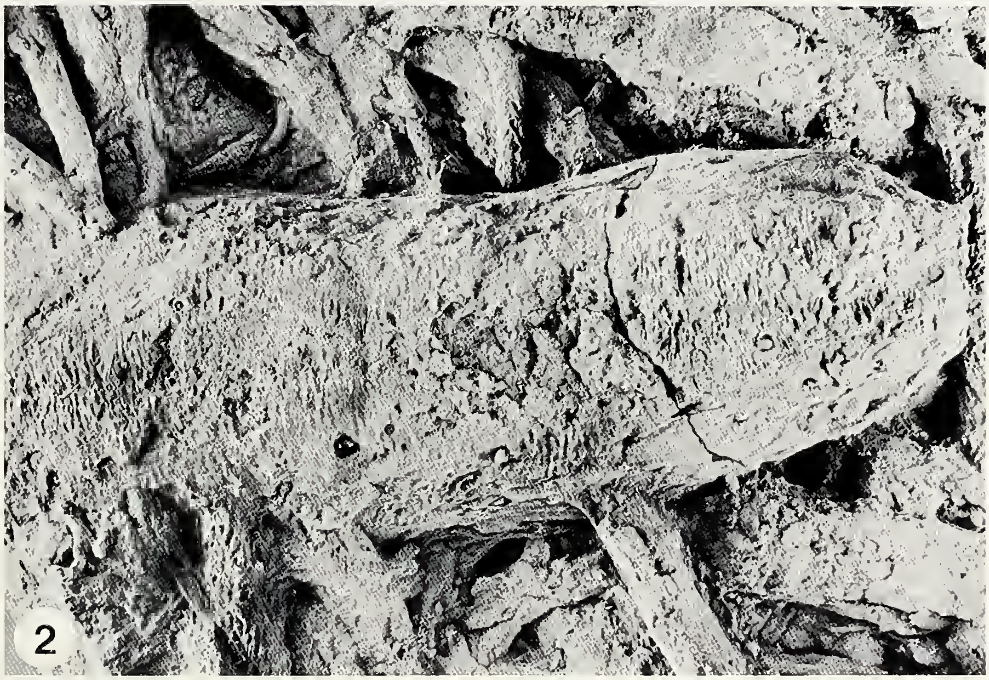


Figure 5 1) Undersurface of Tichenor Limestone displaying elongate burrow(?)fillings. Note intersection and bifurcation of burrows (arrow). Cazenovia Creek, Spring Brook, Erie County, New York. 2) undersurface of basal bed of Hungry Hollow Limestone showing burrow casts of various sizes; note v-shaped scratch marks resembling *Cruziana*; large burrow is similar in size and shape to furrows on base of Tichenor; Hungry Hollow, Ausable River, Arkona, Lambton County, Ontario; X 0.5.

Maynard and Pryor, 1980, p. 25-26). Alternately, fossils such as brachiopod shells, bryozoans, and crinoids may occur in isolated, widely-spaced patches along single bedding planes. This pattern probably reflects an originally patchy distribution of the organisms (e.g., crinoid "colonies," bryozoan thickets, brachiopod clumps). In some examples, however, fossils are broken, aligned and/or exhibit preferred convex-upward orientations indicating some degree of post-mortem disturbance.

Occasionally, elongate ribbon-like stringers of aligned and sometimes edgewise-oriented shells occur along bedding planes in Hamilton shales (Fig. 6; see Linsley, 1972, for discussion of these strips). Lineation and imbrication of shells suggest local movement and alignment of debris by currents. Some shell stringers closely resemble shell-filled furrows described from Recent shelf sediments (Reineck and Singh, 1972; Flood, 1983).

Regardless, shells on the tops of such beds are predominantly oriented convex upward (Fig. 6). Thinner shell beds typically occur as pavements of preferentially convex-up brachiopod valves and elongate, sometimes aligned, fossils (e.g., bryozoans and crinoid columnals).

Most Hamilton shell beds contain a mixture of articulated and disarticulated skeletal materials. Fragile fossils, such as bryozoans and crinoids, are fragmented; however, most brachiopods occur as unbroken single valves. Simple shell beds only rarely contain abraded, corroded, or roll-faceted fossils.

Of particular significance is the occurrence of well preserved multi-element skeletons within certain Hamilton shell beds. Complete trilobites and crinoid and blastoid crowns, both on and within shelly lenses, indicate rapid entombment; these commonly occur associated with debris-rich stringers or laminae of mud-silt matrix within shell-rich beds (for further discussion see Meyer, 1971; Liddell, 1975; Brett and Eckert, 1982). The fact that such fossils may occur at the *base* of shell beds, or may even extend through the beds, implies that the entire sediment mass was reworked and deposited permanently, all within a very short interval of time, thereby preventing disarticulation of multi-element skeletons.

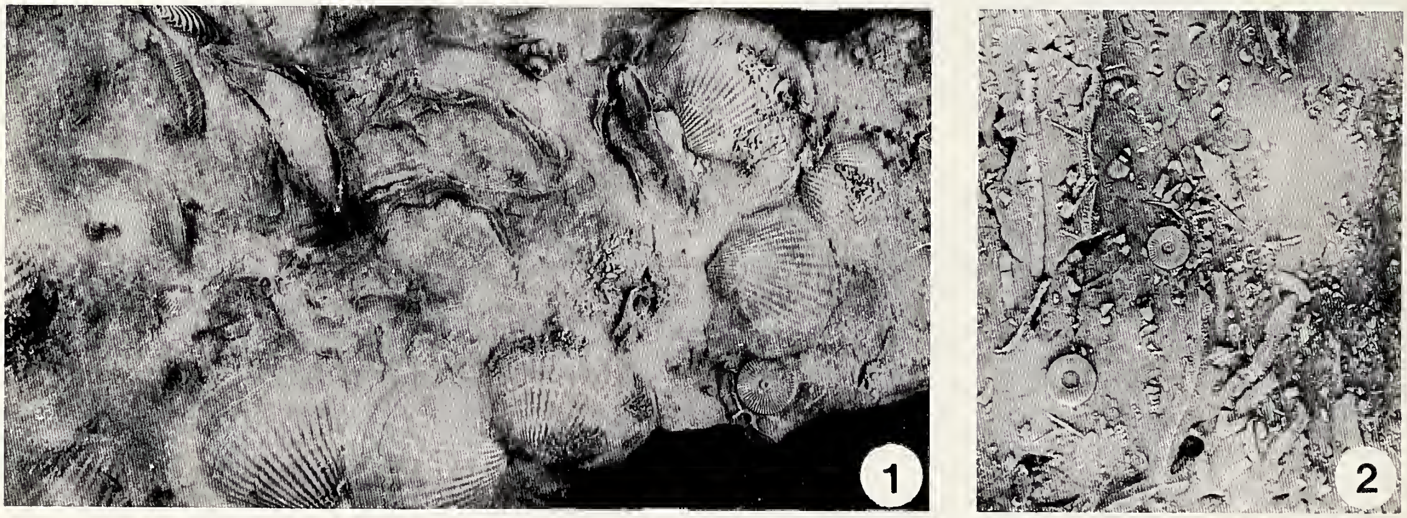


Figure 6 **Parautochthonous coquinites.**
1) Upper surface of lenticular shelly pavement composed mainly of articulated shells of *Tropidoleptus carinatus* Conrad. Note predominant convex upward orientation of *Tropidoleptus* shells; certain specimens also are imbricated. Also note abundant encrusting epibionts, including the cyclostome bryozoan, *Hederella, Aulocystis* corals, and cornulites worm tubes; certain auloporid colonies extend from shells onto adjacent sediment or other shells indicating that encrustation post-dates formation of the shell pavement, i. e., that *Tropidoleptus* shells lay exposed on the seafloor following storm disturbance. Kashong Shale; Bowen Creek; Alexander, Genesee County, New York; X 1.0.
2) Upper surface of shelly coquinite bed showing alignment of thin, bladelike bryozoan fragments (*Taeniopora, Sulcoretopora*) and crinoid columnals. Lower Kashong shale; White Creek, Genesee County, New York; X 1.0.

Hamilton shell beds normally have sharp, flat bases. However, the bottoms of certain shell-rich layers are furrowed with markings which resemble groove or tool mark casts. A few examples of Hamilton "gutter casts" (sensu Aigner, Hagdorn and Mundles, 1978) have been previously documented (Brett, Dick and Baird, in press); these shell-filled prods, like similarly structured furrows in the Tichenor (see above discussion), probably reflect scour-enlarged burrows formed in cohesive muds (Fig. 7).

Hamilton shell layers commonly display chaotic, sometimes imbricated, arrangements of fossils within mudstone matrices.

Size sorting and graded bedding are poorly developed in most Hamilton shell beds, probably because of a relatively limited range of particle sizes available in shell-rich offshore muddy sea floors. However, there are notable exceptions to this generalization. Debris layers emanating from the Staghorn

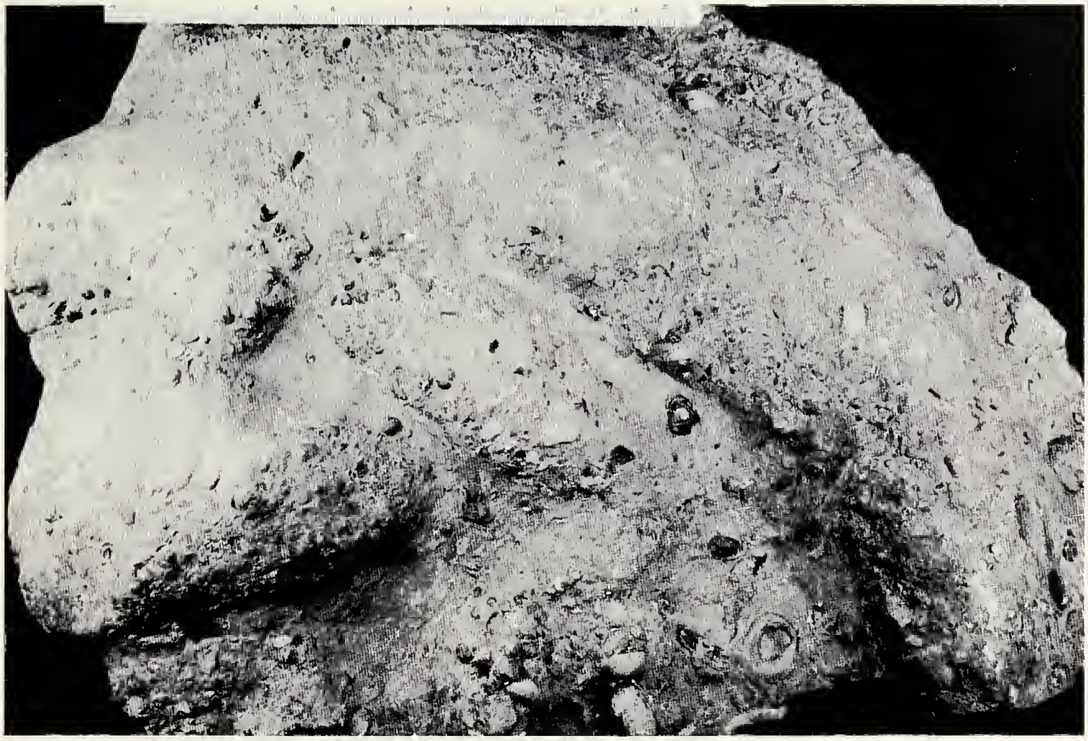


Figure 7 Undersurface of concretionary shell bed showing prominent furrows (gutter casts?) filled with small brachiopods, mainly *Ambocoelia*, trilobite tergites and bivalve shells. Scale in centimeters.

coral biostrome (Fig. 8; see below) exhibit grading with basal lag deposits of rugose corals and fossil fragments passing upward into laminated siltstones. Similarly, beds from within the upper Kashong Shale (Moscow Formation) in western New York exhibit a basal shell pavement which is overlain by laminated calcisiltites (Fig. 9). These beds closely resemble shell-silt couplets attributed to storm deposition by numerous workers (Bowen and others, 1974; Kreisa, 1981; Kreisa and Bambach, 1982; Aigner, 1979, 1982, 1985; Brett, 1983).

Interpretation of Shell Beds

Most Hamilton beds can be classified as parautochthonous coquinites (*sensu* Aigner, Hagdorn and Mundlos, 1978). Although some faunal elements may be restricted to shell beds because of hydraulic sorting or distinctive interstitial diagenesis, such beds usually contain fossil assemblages congruent with those in underlying sediments. This, together with biostratinomic evidence, cited previously, indicates that these fossils have not been transported a significant distance; indeed, in a few layers, some rudiment of the pre-storm community mosaic may have been preserved. In the Staghorn coral debris layers, it is evident that corals and coral debris were transported, at most, a few meters to tens of meters from its source. Whether or not a significant percentage of skeletal debris was reworked from underlying beds remains an open question. Typical shell beds, however, exhibit little evidence for erosion of older sediment. Intraclasts are characteristically uncommon, but mud-filled shells and even closed shells with reoriented geopetal fillings have been observed. In certain beds, sheltering of mud beneath shells indicates that interstitial muds may have been partially winnowed away.

Tractional concentration of shells (including those of organisms still living at the time) into windrows, probably resulted from storm-wave impingement on a muddy sea floor. Storm surge back-flow presumably produced density currents which scoured the seafloor and accumulated debris stringers before silt and mud suspension-burial. Hence, most Hamilton simple shell beds can be interpreted as tempestites (sensu Aigner, 1982; Seilacher, 1982).

Certain of the thinner Hamilton shelly beds, however, cannot be simply interpreted as erosion lag deposits, such as postulated by Bowen and others (1974) for Upper Devonian coquinites; rather they show stratinomic evidence for *in situ* storm reworking and mud burial of skeletal material following intervals of low sedimentation (cf. Fürsich, 1982). This is most evident in shell beds that occur in otherwise sparsely fossilifer-

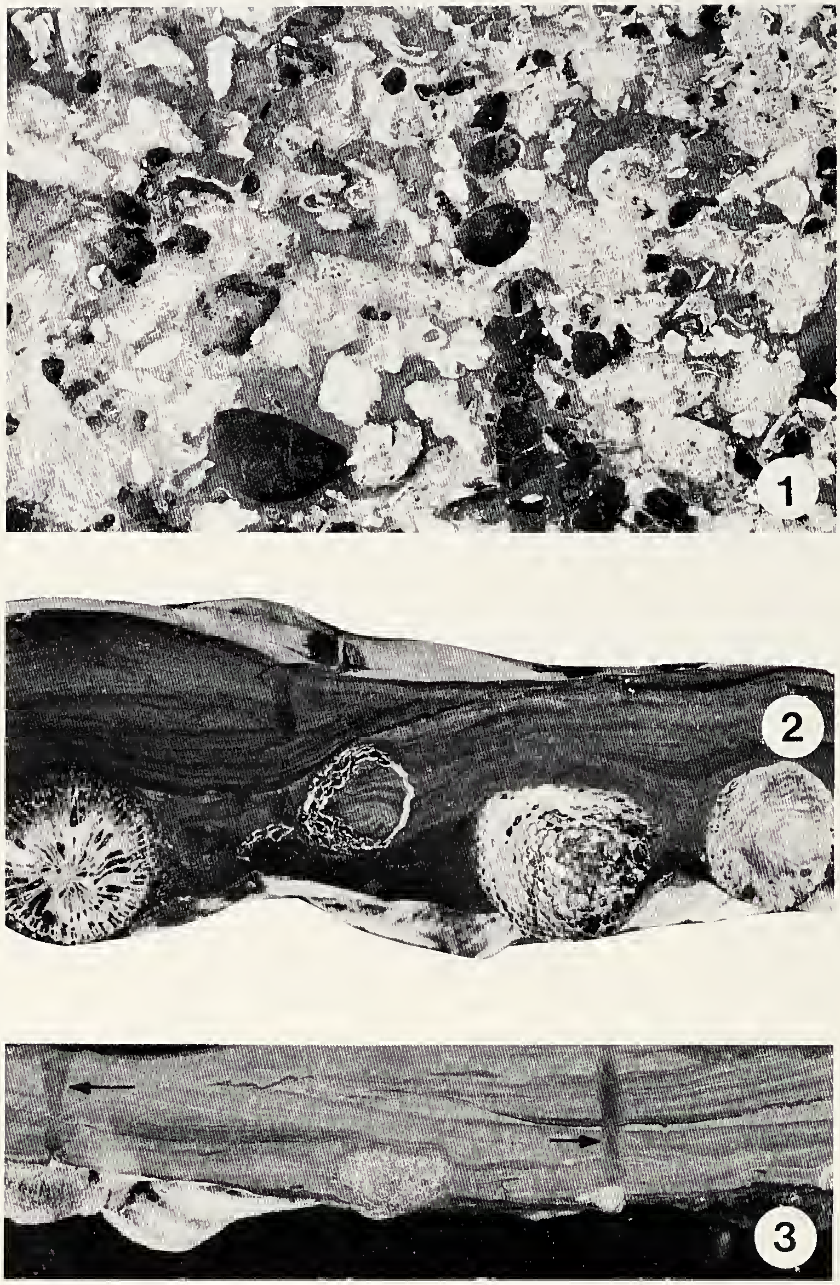


Figure 8 Coral-rich, graded siltstone bed. 1) View of undersurface of slab showing mixture of complete and fragmentary rugose corals, spiriferid brachiopod valves and phosphatic nodules; X 0.25. 2,3)Lateral views of polished slabs; illustrating basal lag of corals overlain by hummocky cross-laminated siltstone; also note vertical shafts of probable escape burrows (arrows). Staghorn coral bed horizon; Otisco Shale Member; Jenny Point, Skaneateles Lake, Onondaga County, New York; both X 0.6.

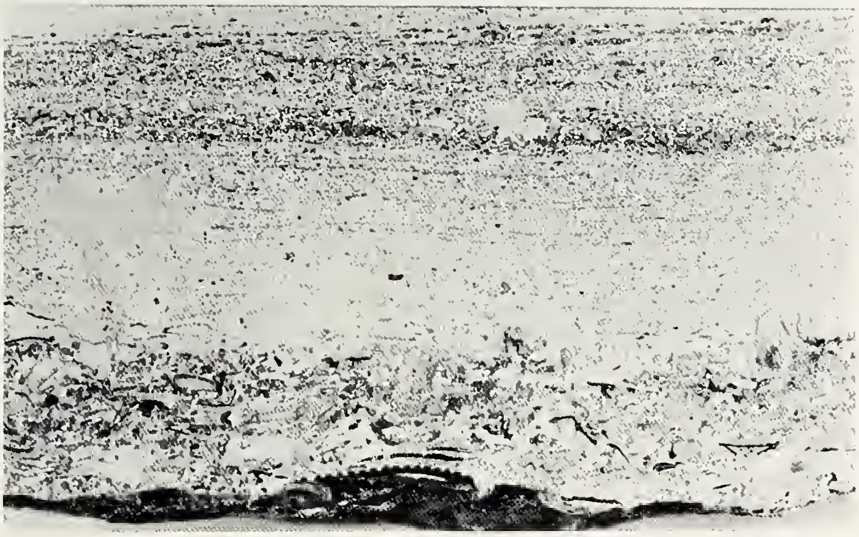


Figure 9 Calcisiltite bed; lateral view of polished slab showing crude grading, with basal layer of coarse shell fragments overlain by finely laminated carbonate silt and pellets; upper Kashong Shale; White Creek, Genesee County, New York; X 1.5.

ous mudstones, notably in the lower Wanakah, Kashong and upper Windom shales. Several of these beds are mappable over distances of tens of kilometers. In these cases, matrices of the shell beds are typically bioturbated and contain few fossils belonging to species that also occur in the surrounding mudstone layers; most forms apparently are restricted to the shell beds.

Excellent preservation of some fossils within many shelly layers, furthermore, strongly argues against significant lateral transport; indeed, certain fossils appear to have been preserved in life positions. Instead, the high diversity and density of fossils within the shell beds may be attributed, at least in part, to "successional" buildup of epifauna during times of low sedimentation rates, either due to regional sediment starvation or to locally altered current patterns. Although some shells may have been reworked by more than one storm event, they were probably buried in their present orientation by a single, "final" storm event prior to renewed sediment deposition. Thus, such beds represent *both* longer term accumulations of skeletal material during calm-water, slow deposition conditions, and instantaneous final entombment within tempestites. Such a model for the genesis of shell beds has previously been proposed by Fürsich (1982).

As noted by Fursich, storm events occur randomly and probably independently of background sedimentation rates. Storm layers formed during times of generally more rapid sedimentation are less distinct than those formed during sediment-starved intervals. Moreover, in many examples, thin storm layers may have been destroyed by bioturbation in aerobic muddy environments. Thus, shelly tempestites in the Hamilton were best preserved at the end of intervals of slow sedimentation.

Shell beds commonly weather out in relief on outcrop surfaces indicating greater resistance than background sediments. Typically, shells occur in a mud matrix which has been partially indurated by selective precipitation of microspar carbonate cements. Moreover, some shell beds display aureoles of cemented background sediments adhering to their undersurface (under beds, sensu Aigner, 1982) or to their upper surfaces, or both. A related phenomenon, observed commonly in the ambocoeliid biofacies (e.g., Ledyard, middle Wanakah and Windom shales), consists of discrete concretions developed symmetrically or asymmetrically along shelly beds. In such instances shell layers pass through the concretions.

In summary, many simple shell beds from the Hamilton Group display features indicative of episodic reworking and burial. These include:a) scoured bases, b) grain-size grading (rare), c) starved ripples, d) selective fossil orientation, e) intact preservation of multi-element skeletons, and f) evidence for disinterment of fossils from underlying beds. In many examples, good preservation of fossils and general resemblance of shell bed faunas to background fossil assemblages indicate *in situ* reworking and burial of skeletal debris.

Laminated Siltstone and Calcisiltite Beds: Storm Silt Layers

Interbeds of laminated terrigenous or carbonate siltstones are relatively uncommon within the Hamilton Group; nevertheless, a few examples may be noted. Finely laminated terrigenous siltstone beds are unknown in the western New York Hamilton but become increasingly common in west-central New York. Here they occur primarily in medium to dark-gray shales and less frequently in lighter gray, bioturbated, silty mudrocks. The beds are typically 0.5 to 2 cm thick and contain fine planar stratification or cross lamination.

The thickness and frequency of occurrence of these siltstone layers increase upward in several upward-coarsening sequences in the Hamilton Group. For example, siltstones appear with increasing frequency toward the top of the Levanna Shale interval, exposed in the Owasco Lake to Syracuse meridians. This trend parallels an upward black to dark-gray shale change and a corresponding increase in fossil content and faunal diversity.

Basal surfaces of the siltstone beds are sharply defined and often show current lineations, including groove casts. Tops of layers may be slightly undulatory or rippled or may appear to grade into the overlying mudstones. Thicker siltstones, such as those associated with the Staghorn Coral bed debris layers (Fig. 8), display hummocky cross lamination and vertical escape burrows (fugichnia). As noted in the previous section, some of these beds have a basal, coarse lag zone of fossil coral debris (Fig. 8.1, 8.2).

In western New York, calcisiltites occur in certain units, particularly in the Deep Run and upper Kashong shales. These (5-10 cm) layers are composed of laminated fine skeletal debris and pellets and display features analogous to the terrigenous siltstones. These features include sharp, erosional bases with scour marks and fossil lags, internal planar to hummocky cross lamination, probable escape traces and burrowed, undulatory top surfaces (Fig. 9). Calcisiltite beds are apparently restricted to mudstones which contain a moderately diverse, *Tropidoleptus*-dominated fauna. Hamilton calcisiltites closely resemble allodapic carbonates described by Brett (1983) from the Silurian Rochester Shale in western New York.

Interpretation of Siltstone — Calcisiltite Layers

Laminated siltstones and calcisiltite layers, occurring as isolated beds within clay, shale, and mudstone sequences, are interpreted as distal storm silt layers analogous to those observed by Reineck and Singh (1972) and Aigner and Reineck (1982) in the North Sea. Indications that these beds were deposited during episodic storm current activity include a)scour features at bed bases, b)vaguely developed grain-size grading, c)internal planar to cross lamination, d)escape traces, and, in some instances, e)hummocky stratification. The last, seen only in thick siltstones and calcisiltites, suggests wave traction-influenced deposition in shallow water, probably 20 m or less (Dott and Bourgeois, 1982).

It is notable that the thicker (2 to 10 cm) siltstone or calcisiltite bands are found in association with strongly bioturbated gray mudstone facies. Thin siltstones, in contrast, are seen only in dark gray to black shale sequences, being more common and thicker in the former. Together, these observations suggest that bioturbation was responsible for the destruction of thinner silty beds in most aerobic Hamilton facies. Only in anoxic to dysaerobic bottom settings were thinner silty layers preserved. This hypothesis is further supported by evidence from laterally traced siltstone bands. Thin laminated siltstones in the lower Otisco Shale grade laterally from non-burrowed beds to more obscure, burrow-riddled layers along a single exposure on Skaneateles Lake. Where the beds thin to a critical limit, they become completely reworked into the synjacent sediment and, therefore, are not easily traced into more distal facies. Similarly, the Mack Creek turbidite described by Baird (1981) becomes totally disrupted and reworked by burrowing as it is traced southeastward from darker gray basin margin shales to upslope medium-gray mudstones in outcrops along the eastern shore of Cayuga Lake. These and similar examples indicate a profound biogenic control on the preservation of such siltstone beds.

The rarity of storm silt layers in the Hamilton Group contrasts markedly with their abundance in previously reported sequences of Ordovician (Kreisa, 1981) and Silurian (Brett, 1983) age. This change appears to correlate with the rise of organisms producing large-scale deposit feeding lebensspuren in the Late Silurian-Early Devonian, most notably with the greatly increased abundance of the trace fossil *Zoophycos*. Several authors have stressed the importance of evolution of deep-burrowing organisms in controlling changing patterns of stratification through time (Thayer, 1981; Sepkoski, 1982; Larson and Rhoads, 1983). The reduction in preservation of storm silt beds in Devonian as compared with earlier Paleozoic sequences may provide an example of this phenomenon; this being, in part, analogous to the disappearance of abundant shallow shelf shale-pebble conglomerates by medial Ordovician time, as documented by Sepkoski (1982). Furthermore, Larson and Rhoads (1983) noted an increase in the mean thickness of sedimentary beds from Ordovician to Devonian times which may be a reflection of the same evolutionary event.

Massive *Zoophycos*-swirled silty or calcareous mudrocks are common within the Hamilton Group and other Devonian (e.g., Esopus Shale; see Miller and Rhemer, 1982), as well as Carboniferous (e.g., Borden Group; Ausich, Kammer and Lane, 1979) stratigraphic intervals. This facies, however, is conspicuously uncommon in Silurian and older rock units. We postulate that the *Zoophycos* mudrock facies corresponds di-

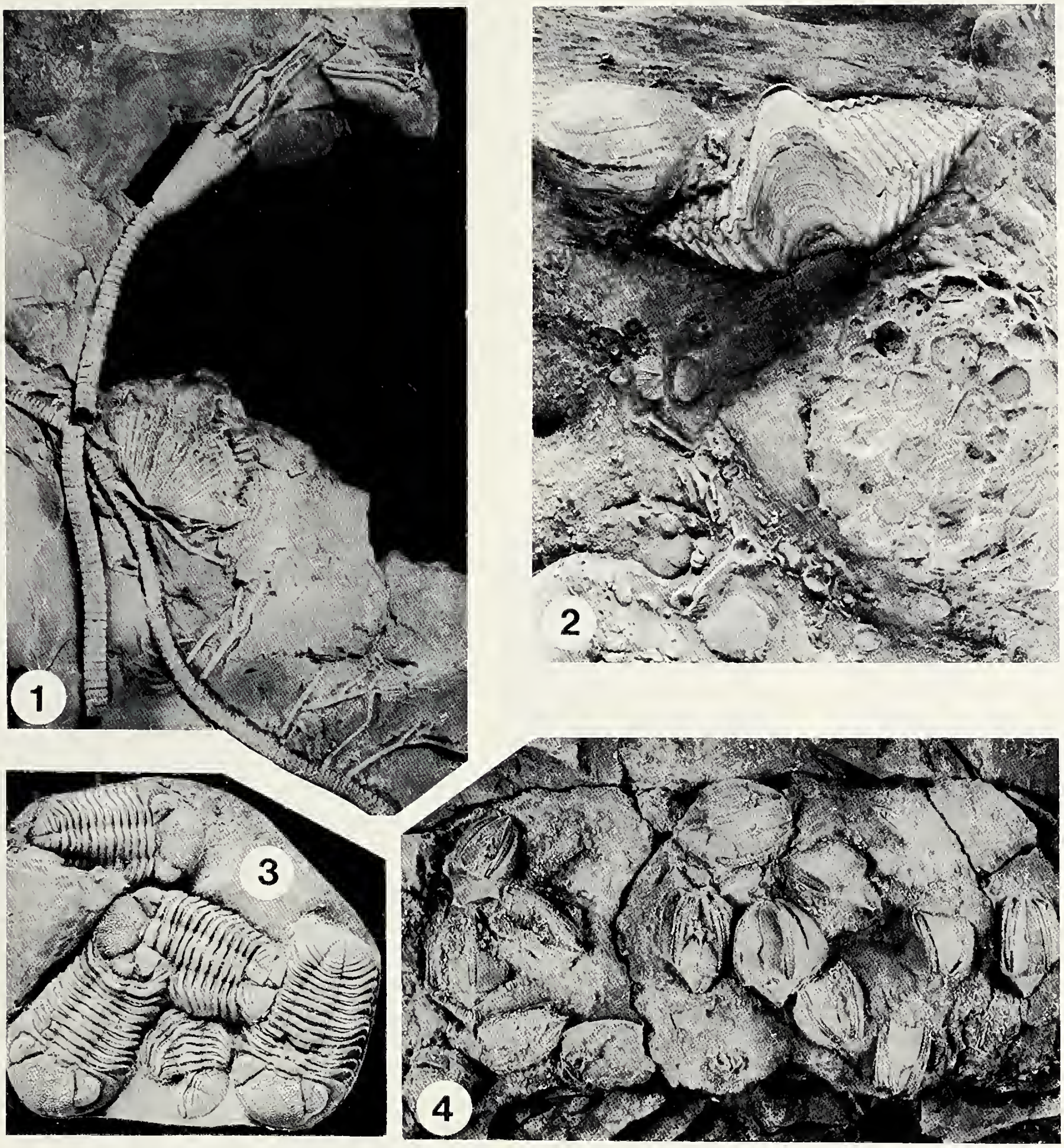


Figure 10 Smothered layer fossil assemblages. 1) Complete specimens of crinoid, *Bactrocrinites*, including cirrus bearing dististele and crown; upper Windom Shale (Taunton beds); Menteth Gully, Canandaigua Lake, Ontario County, New York; X 1. 0. 2) Slab showing spiriferid brachiopod, *Mediospirifer*, and coral, *Pleurodictyum*. Bay View Bed, Windom Shale; White Creek, Genesee County, New York; X 1.0. 3) Cluster of complete *Phacops rana* trilobites; Smoke Creek bed, Windom Shale; Penn Dixie quarry, Bay View, Erie County, New York; X 1.0. 4) Elongate furrow-like structure filled with complete thecae, brachioles and columns of the blastoid *Hyperoblastus goldringae*; lower Wanakah Shale; Cazenovia Creek, Spring Brook, Erie County, New York; 1.5.

rectly to the heterolithic shale-siltstone or shale-calcisiltite facies typical of early Paleozoic midshelf sequences. In cyclic facies sequences of the Hamilton Group this facies appears in positions approximately analogous to the heterolithic calcisiltite-ribbon rock-shale layers of the Silurian Rochester Shale (Brett, 1983), the Ordovician Martinsburg Shale (Kreisa, 1981), and other older units. We infer that heterolithic sequences, thus, are displaced progressively basinward through the Middle Paleozoic such that by the late Paleozoic they occur almost exclusively in anaerobic and dysaerobic settings.

Burial Horizons: Storm Mud Layers

Hamilton strata are famous for exceptionally well-preserved fossils; however, many of these are derived from a relatively small number of thin, though regionally widespread beds, often only a few centimeters thick (Fig. 10). For example, a majority of the complete crinoids figured by Goldring (1923) from the "Moscow at Bristol or Vincent," were apparently derived from a single bed within the upper Windom Shale; our recent studies indicate that this layer is traceable throughout the Bristol and Canandaigua valleys (Fig. 10.1).

Another type of well-preserved fossil bed is represented by the famous Hamilton trilobite beds; i.e., the Wanakah beds (Ludlowville Fm.) of Grabau (1898-1899) and the Windom Smoke Creek bed (Moscow Fm.) described by Brett and Baird (1982) and Baird and Brett (1983). Over the years, literally thousands of complete specimens of the trilobites *Phacops rana* and *Greenops boothi* have been derived from these 20-30 cm thick argillaceous limestone beds (Fig. 10). These occurrences contrast with a general rarity of complete specimens throughout the bulk of synjacent mudstone strata. Most striking is the lateral persistence of these and similar beds; whole trilobites have been obtained from virtually all outcrops of the Wanakah beds (spanning some 90 km from the Lake Erie shore into the Genesee River Valley) and the Windom Smoke Creek bed (mappable from the Lake Erie shore approximately 120 km east into the Canandaigua Lake region). Complete trilobites and intact trilobite exuviae (i.e., moult ensembles) occasionally occur in dense monospecific aggregations. Such clustered occurrences appear to be restricted to a particular, rather limited range of lithofacies, which are intermittently represented by resistant, argillaceous limestone beds, as described herein. The significance and interpretation of the clustered trilobite phenomenon is discussed elsewhere (Speyer and Brett, 1982, 1985).

Beds containing abundant enrolled trilobites have been found associated with concretionary, pyrite-rich zones of the Ledyard and Wanakah Shale members (Ludlowville Formation). Such layers, like the more famous trilobite beds, are laterally mappable and maintain their sedimentary and taphonomic character across broad geographic regions (Speyer and Brett, 1984). Perhaps the most remarkable unit is a 10-12 cm thick argillaceous limestone in the lower Centerfield Member (Browns Creek bed; Baird and others, in prep.) which yields exceptionally well-preserved proetid trilobites. Many of these display a flexure of the exoskeleton wherein the cephalon is oriented subhorizontally and the thoracopygidium is directed diagonally into the sediment (Fig. 11.3) indicating a posture similar to that noted in *Stenopilus pronus* from the Middle Ordovician (Stitt, 1976). In the Hamilton Group this unique mode of orientation, possibly a life or defensive posture, has presently been found only in this single bed (Speyer and Brett, 1984). Nonetheless, these reflexed proetids and other well-preserved fossils have been obtained from the Browns Creek bed at localities spanning an east-west distance of about 50 km in Genesee and Livingston Counties.

These examples serve to illustrate an important and little recognized phenomenon in the Hamilton Group of western New York. Such layers of well-preserved fossils are believed to reflect episodes of rapid mud burial of the sea floor (see below) and hence delineate "smothered bottom horizons" *sensu* Kesling and Chilman (1975).

Beds containing well-preserved fossils may be subdivided on the basis of associated lithology. Many are relatively obscure bedding planes within seemingly homogeneous shales and are recognizable only from detailed taphonomic study; these are herein called "simple burial horizons." In other cases, smothered bottom assemblages are associated with beds of small pyritic nodules or carbonate concretions and may be termed "diagenetically enhanced burial horizons." These are more readily distinguished from background sediments than are simple smothered layers. Certain well-preserved fossil layers occur on or within beds and lenses of argillaceous limestone; these are considered to be largely diagenetic in origin, reflecting a more pervasive overprint.

Simple Burial Horizons

These are "shale-on-shale" horizons marked by bedding plane accumulations of well-preserved fossils which underlie a layer of rapidly deposited sediment. Strata immediately underlying the smothered bedding plane are commonly fossiliferous; fossils are normally disarticulated and randomly oriented in the mudstone matrix which may display an abundance of simple burrows or spreiten. This material is simply considered to represent background sediment. Occasionally, however, this underlying sediment may contain complete, well-preserved fossils in presumed burrow orientation (e.g., articulated bivalves from the basal Jaycox Shale; see Fig. 11.1, 11.2).

The buried sediment-water interface itself is often marked by a slightly irregular pavement of disarticulated shells and debris which suggests a minor diastem or omission horizon. This "smothered" interface is characterized by well-preserved fragile fossils such as complete, outstretched trilobites, crinoids and blastoids, or the presence of articulated brachiopods or rugose corals preserved in life orientations.

Excellent examples of this type of simple burial horizon are encountered in the Wanakah, Kashong, and Windom Shale members. Clement (1981) described a layer of complete crinoids from the upper Wanakah Shale in Erie County. This bedding plane is marked locally by a minor pavement of brachiopods, including complete and disarticulated individuals of *Spinocyrtia*. Complete crinoids, some still attached by holdfasts, also are present. There is little evidence for selective orientation of the fossils. Brachiopod valves occur in both concave-up and concave-down orientations; there are even specimens which indicate flipping from convex-up to the unstable convex-down position immediately prior to burial, as they have small crinoids attached to the (now inverted) convex surfaces of the shell. Similar horizons from the Otisco and Kashong shale display masses of crinoid columns, some with crowns intact, which have been slightly current aligned. Beds with numerous complete spiriferid brachiopods oriented with interareas parallel to the substrate are common in the upper Windom Shale of the Finger Lakes area (Fig. 10.2). In most cases the key bedding plane is overlain by at least a thin layer of relatively barren shale which is interpreted as the burial layer proper. This matrix may contain portions of fossils extending upward from the buried bedding plane, but is usually structureless.

Simple smothered layers are difficult to locate and trace from one outcrop to the next. Nevertheless, where a deliberate effort has been made, layers can sometimes be locally correlated (Clement, 1981; "*Clarkeocrinus* Bed" in Bristol Valley region, unpubl.data of Brett).

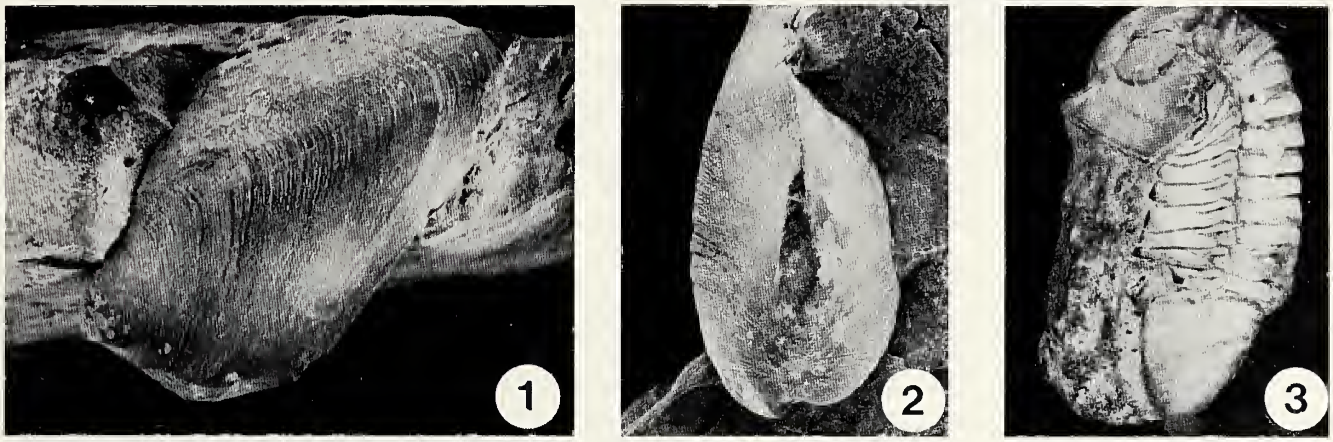


Figure 11 **Preservation of fossils in probable burrow positions. 1,2) Lateral and lower surface views of complete tightly closed specimen of *Modiomorpha* cf. *M. subalata* in apparent life orientation within *Zoophycos*-bioturbated calcareous mudstone; basal bed Jaycox Member; Jaycox Run, north of Geneseo, Livingston County, New York; X 1.0. 3) Specimen of *Dechenella rowi* displaying reflexed configuration; note subhorizontal orientation of cephalon and oblique inclination of thoracopygidium; Browns Creek bed, lower Centerfield Member; railroad cut at intersection of Center Road, Bethany, Genesee County, New York; X 1.5.**

Diagenetically Enhanced Burial Horizons

Pyritic or concretionary nodules may occur in mudstones associated with complete fossils indicative of burial events. The most notable example of this type of occurrence is the prominent main nodule layer of the "Alden pyrite bed," Ledyard Shale Member of the Ludlowville Formation (Fisher, 1951; Dick, 1982). Here, nodules of pyrite are nucleated on well preserved, pyrite-infilled fossils, including enrolled trilobites, articulated bivalves, gastropods, blastoids, and other taxa. The main nodule horizon, which is 5-10 cm thick, occurs within a generally pyrite-rich interval; it apparently marks a buried bedding surface.

Another level of small nodules occurs in Ledyard mudstone about 2 m below the Alden pyrite nodule bed. Here small carbonate concretions sometimes contain complete inverted trilobites and phyllocarid skeletons (Fig. 12). Nodules are approximately the same size as the fossils; where nodules are barren, it is thought that nucleation occurred on unpreservable soft bodied organisms. A number of other less conspicuous thin beds of small pyritic and/or calcareous nodules occur at various levels in the Ludlowville and Moscow Formations and occasionally contain well preserved fossils indicating sudden burial.

Larger carbonate concretions (up to a half-meter in diameter) are normally confined to narrow zones, some of which have been correlated for several kilometers (e.g., Spring Brook and Walden Cliffs concretion beds of the Wanakah Shale; Kloc, 1983). Although many of these concretions are com-

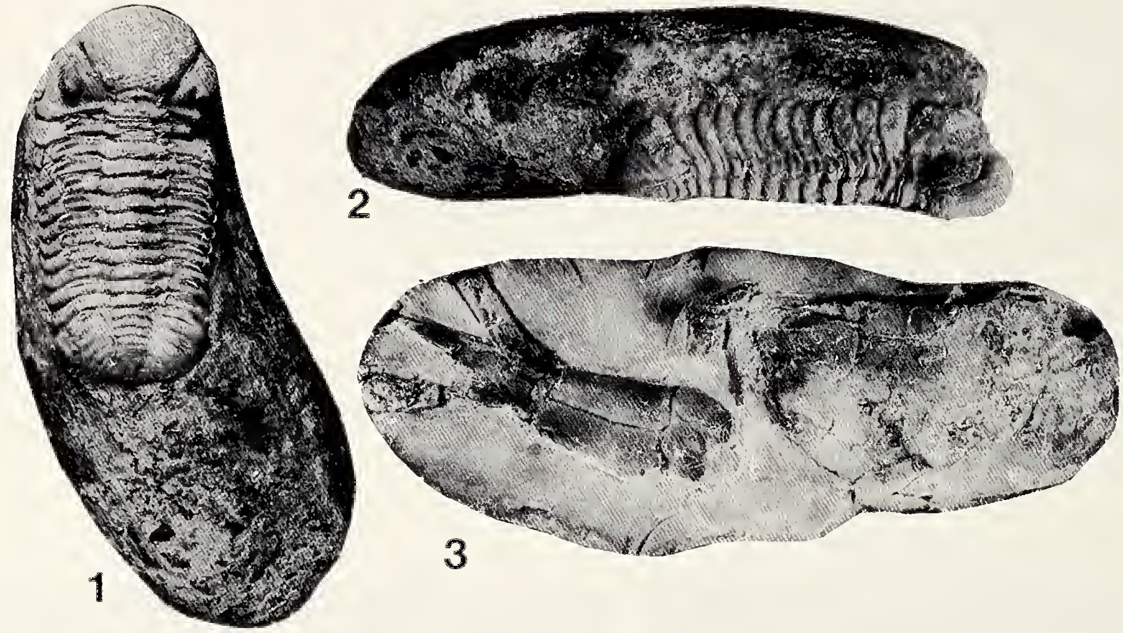


Figure 12 Nodular fossil occurrences: 1,2) Bottom and lateral views. Trilobite *Phacops rana* complete exoskeleton partially embedded in calcareous nodule; note that exoskeleton is inverted (dorsal shield downward) and at the base of the nodule 0.75; both X 1. 0. 3) Nearly complete carapace and abdomen of phyllocarid *Rhinocaris,* enclosed within a carbonate nodule. Both specimens upper Ledyard Shale; Cazenovia Creek, Elma, Erie County, New York; specimens courtesty of G.J. Kloc, Buffalo, New York.

posed of sparsely fossiliferous argillaceous micrites, some contain well preserved and uncompacted fragile fossils. For example, the Spring Brook concretionary bed yields enrolled, pyritized trilobites at many outcrops, both within the concretions and in laterally equivalent nonconcretionary shale. Horizons of discrete large carbonate concretions in Hamilton shales appear to grade laterally, within single outcrops, into tabular argillaceous carbonate beds. These bands are often subconcretionary, displaying minor local pinching and swelling. Nonetheless, they, like concretionary beds, are often widely mappable and tend to maintain remarkably consistent average thicknesses over great distances.

Most argillaceous limestones in the Ludlowville and Moscow Formations are relatively thin, ranging from 2 to 30 cm in thickness. These limestones have internal fabrics and faunal assemblages which closely resemble those of the subjacent shales. Shells within these beds, as in concretions, typically show little compaction as compared with synjacent shales in which fossils tend to be highly compressed. Such concretionary, argillaceous limestones, unlike more prominent skeletal limestones (e.g., middle Centerfield, Tichenor, or Portland Point beds) do not appear to be directly related to cyclic facies sequences; rather, they appear to occur independently of facies cycles. Nonetheless, they are more common in certain facies than others, being most characteristic of dark to medium-gray, slightly calcareous mudstones associated with ambocoeliid, chonetid, brachiopod and/or the stereolasmatid coral biofacies. Concretionary limestones are uncommon in black shales and are virtually absent in thicker silty mudstones. Pyrite is normally uncommon in such argillaceous limestone beds, but sometimes occurs as thin linings on endichnial thread burrow traces.

Many of these argillaceous beds exhibit excellent preservation of fossils, including complete trilobites (Fig. 10.3) and crinoids, along discrete layers within the limestone bands; these occurrences presumably indicate the buried substrate surface. Indeed, the widely known "trilobite beds" of the Hamilton Group (e.g., Wanakah trilobite beds and Windom Smoke Creek bed) occur in such argillaceous carbonate units. Sequences within the limestone beds resemble those described above for simple smothered bottom layers, including sparsely fossiliferous background sediments, middle layers of exceptionally well preserved fossils and less fossiliferous burial material (Speyer, 1983). As noted above, certain of these calcareous beds consistently exhibit unique preservational features, of which the reflexed proetids of the Browns Creek bed are but one excellent example. The associations of these anomalous concretionary carbonates with well-preserved fossils, the reduction or absence of compaction within them, and their lateral gradation into discrete concretions strongly suggest that these limestones reflect selective early diagenetic precipitation of carbonate cements within mud, in response to burial events.

Interpretation of Burial Layers

Because organisms with multi-element skeletons, such as trilobites and crinoids, disarticulate very rapidly following death (see Meyer, 1971; Liddell, 1975), their complete preservation is a rare event, normally requiring very rapid burial. Hence the taphonomy of trilobite and crinoid beds, such as those reported herein, provides unequivocal evidence for sudden death and burial of benthic communities. Moreover, the occurrence of unique and nearly identical biostratinomic features within such beds, over larger areas, requires a specific combination of depositional circumstances. Nearly simultaneous burial apparently occurred over large areas (probably several hundred square kilometers) of relatively uniform sea floor environment.

The repeated occurrence of widespread, persistent layers in western New York strongly suggests a parallelism between the present outcrop belt and Devonian facies tracts through time; the shelf environment was elongate east-west, paralleling the northern margin of the Appalachian Basin and the present day outcrop belt (see Brett, 1983, for further discussion). This would explain the similarity of sea floor conditions at the time of burial along the outcrop transect. The wide lateral extent of many buried layers across three or four counties of western New York, strongly suggests a powerful and large scale agent for sediment resuspension across this region. Brongersma-Sanders (1957) maintained that catastrophic events similar to contemporary hurricanes resulted in paleontologically significant mass mortality phenomena. Studies by Engle (1948) and McKnight (1969) document heavy mortality within recent marine benthic communities buried by muds following storm episodes. Suspension clouds have been observed near recent shelf bottoms, following severe storms (Rudolfo et *al*., 1971; Reineck and Singh, 1972; Aigner and Reineck, 1982). These clouds of sediment are transported basinward where, augmented by local resuspension (Drake, 1976), they are deposited on the sea floor as mud layers.

Kranz (1974a, b) has demonstrated the "escape potential" of many vagrant organisms that in the Devonian included trilobites, mollusks, and probably stelleroids, pointing up the significance of thickness of the burial layer in preserving the smothered bottom assemblages. It also is probable that most 'smothered bottom' phenomena were of minor effect and were, consequently, obscured by bioturbation and recolonization. Thus, the beds discussed herein probably represent only the most severe among many burial events. It is perhaps not surprising, then, that these events resulted in the burial of large tracts of sea floor across much of western New York. Such events preserve very rare paleontologic snapshots of the ancient sea floor (*sensu* Schindel, 1980) and provide fossil assemblages which closely represent the original biocoenoses.

The relative frequency of these burial events can be estimated by making several assumptions. The Moscow Formation, probably representing about a fourth of Hamilton time or about 1 to 2 million years, contains at least ten widespread mass mortality horizons suggesting a frequency of about one per 100,000 to 200,000 years. Probably many more, less obvious horizons (i.e., simple burial beds) are as yet undiscovered and may remain so because of having been "erased" by later events or burrowing; thus, the above estimates are almost certainly overinflated. A frequency of one layer per 10,000 years probably is more reasonable. Recently, a detailed study revealed over ten distinct burial horizons within a 1.5 m interval of the upper Windom Shale ("Taunton beds") of the Genesee Valley (Speyer, unpublished data). These data imply an even greater frequency of localized storm events.

Regionally deposited storm resuspension layers may provide isochronous surfaces comparable to, but much more ubiquitous than, ash layers, which are ideal for intrabasinal correlations (for discussion of use of bentonites in coenocorrelation see Cisne and Rabe, 1978; Rabe and Cisne, 1980). Where burial layers have been diagenetically "overprinted" they may be readily detected and mapped as carbonate concretions and/or layers.

As noted above, some of the smothered bottom assemblages are associated with selective syndiagenetic events, including pyrite or carbonate concretion formation. Similar phenomena are associated with episodic mud burial elsewhere, including several famous fish-bearing concretionary levels (Weeks, 1953, 1957; Berner, 1968; Sass and Kolodny, 1972). Waage (1964) reported a well-preserved, virtually intact molluscan fauna from carbonate concretions of the Late Cretaceous Fox Hills Formation of South Dakota. He suggested that the recurrent concretionary horizons actually reflect multiple mass mortality events wherein organic remains have been preserved as concretion nuclei via diagenetic precipitation of calcium carbonate. Although Waage did not expound upon the geochemical processes, he implied a connection with rapid burial and incipient anaerobic decay of organic matter. Presumably, organic decomposition resulted in a substantial alteration of interstitial chemistry and, therefore, acted as a catalyst in initiating calcium carbonate precipitation (Tarr, 1921; Burt, 1932).

Weeks (1953, 1957) suggested a similar mechanism for carbonate precipitation, citing spectacular Cretaceous concretions which contain three-dimensionally preserved fish remains. He reasoned that the decomposition of the fish produced isolated "chemical microenvironments" around which calcium carbonate precipitated. Berner (1968), and later Zangerl and others (1969), demonstrated that, as had been predicted, a substantial amount of ammonia appears in ambient interstitial sea water following prolonged decomposition (approx.135 days) of organic material (i.e.,fish). The production and release of ammonia, likewise, coincided with a pronounced decrease in dissolved calcium (Ca^{+2}) and a concentration of calcium soaps (adipocere) about each fish carcass. Berner (1971) later outlined the two most important chemical systems involved in the mobilization of dissolved ions and the diagenetic precipitation of calcium carbonate ($CaCO_3$), bacterial reduction of sulfate (SO_4), and the release of ammonia during deamination as summarized above.

Although a considerable literature treats the *in situ* precipitation of calcium carbonate around organic nuclei, no studies, except that of Waage (1964), have discussed the sedimentologic and paleontologic backdrop against which this geochemical phenomenon is set. We propose that catastrophic burial events during deposition of the Hamilton beds often resulted in widespread mortality which, in turn, initiated the series of diagenetic processes outlined by Berner (1971) (Fig.13). Quite obviously, proper geochemical conditions must be prevalent on the sea floor at the time of storm incursion for diagenetic processes to be operative. Thus, not all burial horizons underwent subsequent diagenetic modification. Diagenetic 'overprinting' of event stratigraphy is apparently most often associated with smothered bottom horizons in slightly dysaerobic, nonsulfidic substrates in which low pH/Eh conditions existed within the burial sediment. In such environments anaerobic decay of freshly buried organic matter may have generated concentrations of ammonia, locally raising pH of the sediment and favoring precipitation of interstitial carbonate cements.

We do not mean to imply that all carbonate precipitation required to form a given concretion or limestone bed resulted from the local disturbance of pH due to organic decay; we merely suggest that this process *initiated* the precipitation of carbonate cements which stabilized sediment in an "aureole" around the buried organic matter. Once the grain framework of a sediment layer had been stabilized by incipient cementation, it should have acted as a site of further cement precipitation during subsequent gravitational compaction (see Einsele, 1982; Eder, 1982; Walther, 1982). Surrounding muddy sediments, once compacted, will tend to lose most of their dissolved carbonates because of the absence of pore space for cementation (see Einsele, 1982, p.28). We further suggest that this process is most active in muds with initially high concentrations of carbonate material.

Diagenetic enhancement (cf.Eder, 1982; Walther, 1982) of storm deposited strata not only facilitates the recognition of such event horizons in outcrops but also preserves, to some degree, original deposit thickness and sedimentary fabric. Concretionary bodies within shale from the Late Cretaceous Mishash Formation of Israel reflect early diagenetic precipitation of calcium carbonate within interstitial pore space. Such evidence indicates that stratal thickness may be reduced by as much as 80% as a result of postdepositional sediment dewatering and compaction (Sass and Kolodny, 1972). Typically, depositional fabric features (e.g., laminae) are offset or distorted across the shale/concretion boundary such that narrowly spaced laminae throughout shale sequences are expanded within adjacent concretion bodies. In addition, trace fossils are often preserved as a result of diagenesis, whereas, under normal sedimentary conditions (i.e., compaction) these structures would be obliterated by compaction. In many cases individual burrows are accentuated by a pyritic rim presumably precipitated as ambient sea water supplied sulphate anions (SO_4) to oxygen-deficient sediment. Preservation of these structures is further enhanced by interstitial carbonate precipitation and is, therefore, an indirect effect of diagenetic replacement of primary pore space by calcium carbonate (Eder, 1982). Overprinting of background sediments underlying the burial horizon proper is believed to be a function of calcium mobilization and the dimensions of a microchemical aureole which, in turn, are controlled by calcium availability and the amount of decomposable organic material, respectively (Speyer, 1983).

Commonly, background strata included within the limestone beds display numerous, simple burial layers which have been "overprinted" by the diagenetic process associated with the burial and preservation of the topmost smothered horizon. The occurrence of such composite beds (e.g., Murder Creek Bed and Smoke Creek Bed at most exposures) indicates that smothered bottom diagenesis is a complex process which is not initiated by all burial events. The underlying, "overprinted" event horizons are commonly discontinuous and display various degrees of bioturbation. It is not clear why certain burial events initiated the diagenetic process while others did not. It may be that an interstitial geochemical cycle of sorts moderates the diagenetic expression of sedimentary events such that a threshold concentration of calcium and carbonate ions must be reached before precipitation, initiated by episodic burial and organic decomposition, can take place.

Regional correlation of burial layers indicates that bed characters may vary along inferred bathymetric gradients within an oxygen-stratified basin (*sensu* Rhoads and Morse, 1971). The Windom "Smoke Creek Bed," for example, persists as a tabular calcareous bed for some 80 kilometers along strike. Within the Finger Lakes Region (Canandaigua Lake), however, the outcrop belt obliquely crosscuts strike, thereby exposing the onshore-offshore facies spectrum (Baird and Brett, 1981). In this region, the carbonate bed grades laterally into a concretionary horizon which eventually grades into shale east of Canandaigua Lake. Lateral changes in bed character from lenticular to concretionary may be related to a decrease in the amount of calcium carbonate and/or associated organic remains. Furthermore, Dick (1982) and Dick and Brett (1982) provide evidence of mass mortality events, affecting a so-called "dysaerobic" environmental zone; this phenomenon may be responsible for early diagenetic pyritization of buried local faunas. Thus, discrete, locally persistent, pyritic nodule horizons are considered indicative of episodic environmental stress, often in conjunction with rapid storm burial.

The precise chemical interplay between calcium and ferrous iron cations within marine environmental systems is poorly understood (Raiswell, 1976). Nevertheless, field relationships indicate that a coarsely defined continuum of early diagenetic products may be recognized. As noted above, certain beds display lateral changes from tabular or lenticular argillaceous limestone beds to carbonate concretion horizons and, perhaps, finally to horizons with pyritized faunas. Presumably, this gradient represents laterally-successive storm-generated diagenetic facies along a depth related geochemical gradient. In addition to relative amounts of decomposed material, the pyritization process is, most likely, regulated by influx of sulphate ($SO_4^{=}$) through bioturbation from ambient sea water coupled with the mobilization of ferrous iron (Fe^{+2}) into the vicinity of buried microchemical "pockets" (see Dick and Brett, 1982; Brett and others, in press).

PROXIMALITY OF HAMILTON EVENT BEDS: A MODEL

General Model

The degree to which a particular area of sea floor is affected by storm action will depend upon the severity and duration of the storm as well as the depth of the sea. In general, the intensity of storm scour effects will decrease with increasing depth such that a spectrum of deposits will result from any given storm (Figs. 14, 15). Furthermore, a given depositional environment, all other things being equal, will tend to accumulate similar types of storm-generated deposits through time. Thus, tempestites should serve as sensitive indicators of lateral bathymetric change, as well as sea level fluctuations, in any given section.

Idealized models of storm-generated sedimentation on gently sloping sea floors have been presented by Aigner (1982, 1985) and Brett (1983). During severe storms wave base is lowered such that storm rollers impinge directly on the sea floor in a relatively narrow belt (Fig. 14). In these regions the bottom will be scoured, resulting in winnowed lag accumulations of shelly debris ("swell lags" of Brenner and Davies, 1973). Shallow regions that frequently are affected by direct storm scour should be characterized by multiple reworked, amalgamated beds of coarse shelly sediments. Such beds should diverge basinward into a series of thinner shelly lags, each representing a more discrete, stratigraphically-resolved storm deposit. Fine-grained sediment, resuspended by winnowing in more proximal regions, would be transported basinward, following a storm, in the form of silty or muddy suspension clouds. Laminated silty beds should then be deposited proximally and mud

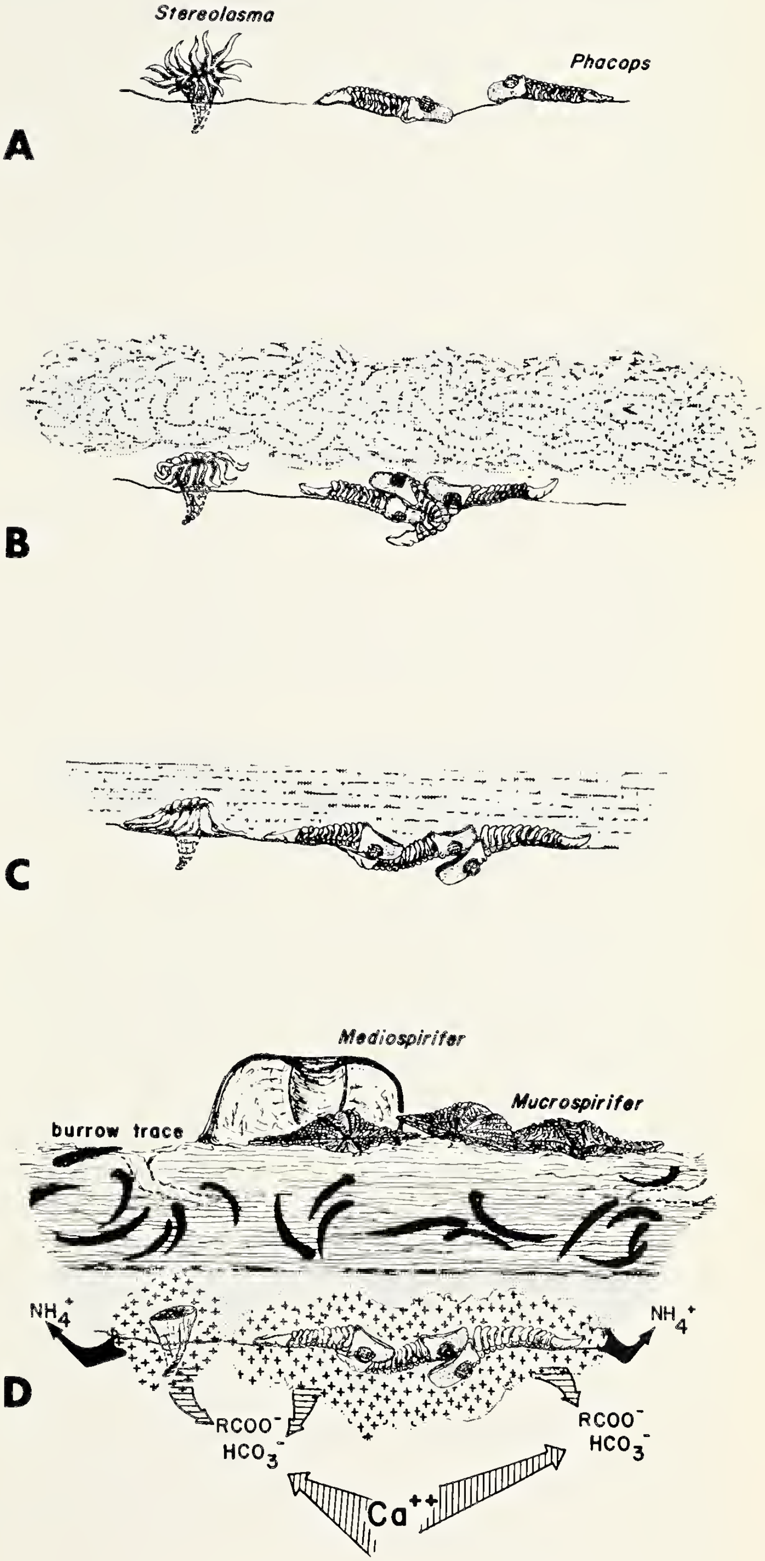


Figure 13 *In situ* burial of benthic faunas beneath storm-generated clouds of sediment. A) Normal background condition; organisms display normal behavioral patterns. B) Sediment cloud is transported from proximal, storm-winnowed facies into deeper water where sediment density eventually overcomes current velocity. C) Deposition of suspended sediment results in mass mortality unless buried organisms can burrow out of mud layer. D) Widespread mortality and incipient decay results in a buildup of various chemical components which ultimately favor early carbonate diagenesis (see text for details).

layers distally. Figure 15 provides a diagrammatic illustration of the predicted proximality spectrum of storm-generated deposits.

To date, few studies of modern storm processes have been presented for areas of mixed carbonate and terrigenous sedimentation. However, Aigner and Reineck (1982) have made a systematic study of storm-generated sediment layers for a graded terrigenous shelf environment in the North Sea. Their findings are in accord with the generalized model provided herein. Cores taken in shoreface and nearshore settings display amalgamated storm sandstone beds, reflecting multiple reworking of sands. Proximal (shallow shelf) environments (5 – 20 m) show an alternation of discrete, laminated storm sand layers and bioturbated sandy silts. Distal settings at depths of 50 to 55 m display an alternation of silty or sandy storm beds and background mud layers. The authors also recognized nonbioturbated mud layers which they interpreted as mud tempestites. Thus, the theoretical model receives strong empirical support from modern shelf environments.

In the context of this model, we interpret Hamilton Group storm beds as follows: first, multiply reworked, coarse lag deposits (e.g., Tichenor Limestone) are considered to represent amalgamated storm beds, the most proximal, or shoreward, storm facies, formed as a winnowed sediment-starved shoal at depths of 5 to 20 m. Sequences of simple shell beds alternating with bioturbated muds, silts, or fine sands comprise the proximal tempestite lithofacies and represent depositional environments generally well below normal wave base, but subject to rare episodic storm wave scour. Sections showing thin calcisiltites or siltstone layers represent medial (transitional; *sensu* Aigner and Reineck, 1982) tempestite lithofacies; these are rare in the Hamilton Group but, as discussed above, may be represented by the strongly bioturbated, silty mudstone facies. This facies reflects areas of intermittent deposition of storm silts, fine carbonate debris and, perhaps, pelletal material, wherein such layers have been largely obliterated and homogenized into background sediments by intense bioturbation.

Finally, smothered bottom horizons reflect the most distal effects of storm action. The deposition of suspended muds occurred rapidly, perhaps due to flocculation effects, such that widespread deep-water regions were simultaneously blanketed. The degree of early diagenetic overprint associated with these mud burial events was apparently related to the geochemistry of the sea floor at the burial site. Where carbonate contents of background sediments were high, rapid burial of organic matter may have resulted in the formation of tabular calcareous mudstone beds. Where available carbonate was more limited, and/or organic matter more patchy, localized concretions tended to form. Similarly, in dysaerobic environments with more reducing, nonsulfidic sediment, rapid burial may have initiated early diagenesis of pyrite as shell fillings, nodules, or crusts, associated with layers or local concentrations of buried organic material.

Because the carbonate content of the sediment, as well as the degree of sediment oxygenation, may have been depth related in the stratified Appalachian Basin, it is suggested that the three types of smothered bottom horizons (i.e., simple layers; carbonate lenses/concretions; nodular pyrite) may be, themselves, sequentially arranged along a proximal-distal axis. Thus, it is believed that simple layers are more common in proximal upslope settings whereas concretionary and pyritic beds record more distal downslope environments.

An Example

A single remarkable outcrop of the Otisco Shale Member of the Ludlowville Formation provides an excellent example of tempestite lithologies and allows for partial testing of the proximality model outlined above. The lower Otisco Shale is exposed nearly continuously along the east shore of the Skaneateles Lake from Jenny Point southward to Staghorn Point (Fig. 16); this outcrop includes the northwestern limit of the well known Staghorn coral bed, thoroughly documented in studies by Smith (1912, 1935) and Oliver (1951). The Staghorn coral bed, a 0.5 – 2 m thick mass of densely packed rugose corals, including *Heliophyllum*, *Heterophrentis*, *Cystiphylloides*, and *Siphonophrentis*, is best developed southward of Staghorn Point where it rests on a siltstone platform; this siltstone bed caps a small upward-coarsening hemicycle, about 3 m thick, which begins in medium-gray concretionary shale and culminates in coarse blocky *Zoophycos* — bioturbated siltstone. Both Smith and Oliver inferred that the coral biostrome was built upon the siltstone platform, the corals initially growing on the firm substrate provided by silty sediment and, subsequently, upon each other.

The platform and the coral beds proper disappear laterally to the north of Staghorn Point; equivalent sediments, intermittently exposed north of these, display a lateral facies sequence associated with the basinward edge of the siltstone platform. Smith (1912, 1935) mapped this sequence and interpreted it as a channel within the siltstone platform. The platform edge rose at least 3 m above a muddy sea floor and was evidently an irregular, cuspate, submarine escarpment which sloped gently to the northwest. The irregularity of the present outcrop along the Skaneateles shoreline has resulted in the exposure of three oblique profiles across this "drop off" from platform edge to the adjacent proximal slope. A relatively steep gradient of approximately 8° – 20° toward the northwest can be directly observed and traced between Jenny and Staghorn Points. At the south end of the section the siltstone platform is still completely developed and is capped by a relatively thin layer of corals, about 20 cm thick as compared with about 2 m of corals at Staghorn Point proper. Northward from this area, within about 500 m, the platform edge appears to have been truncated such that the topmost siltstone and then successively lower levels of the subjacent concretionary mudstones lie in direct contact with a

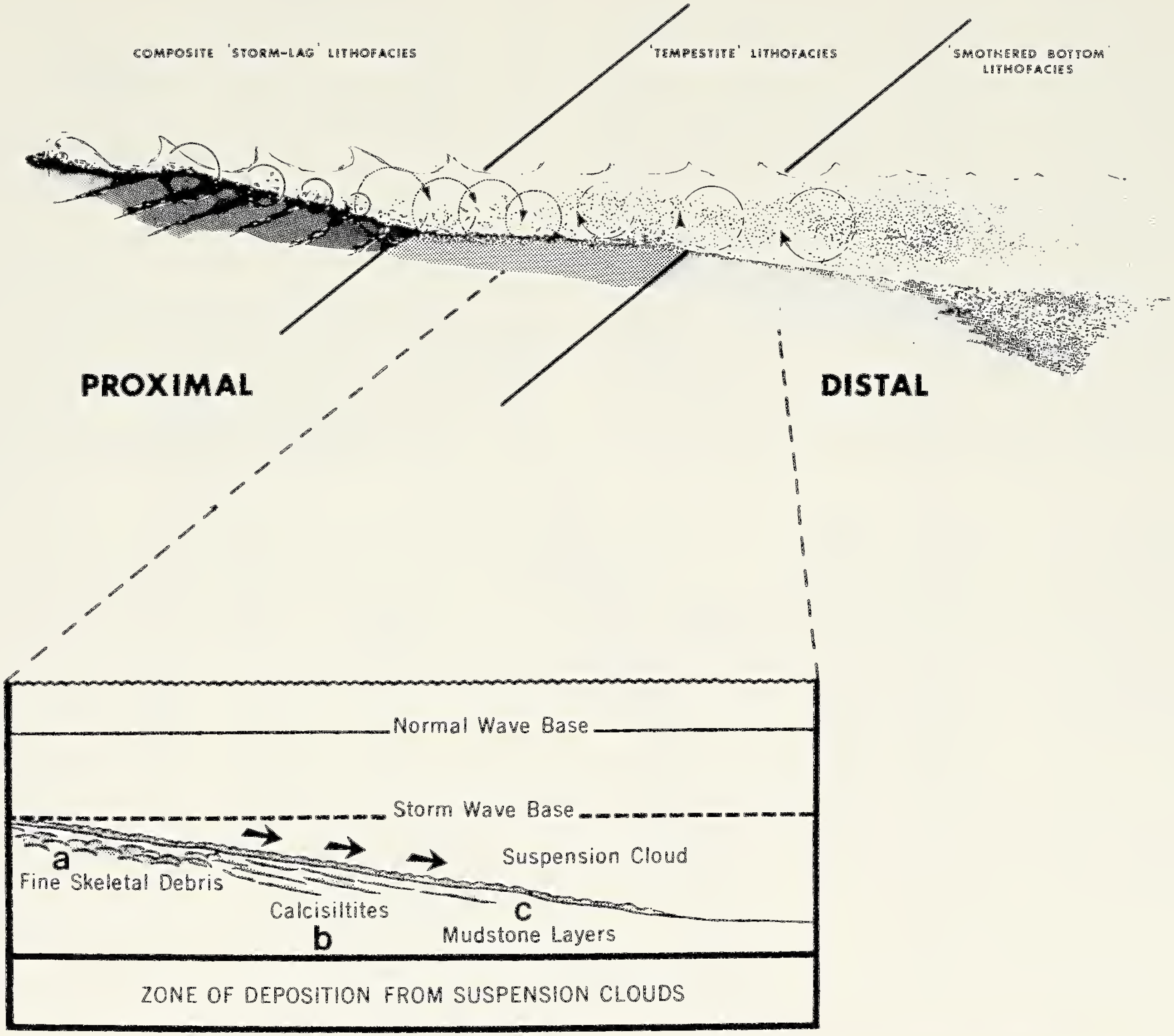


Figure 14 Diagrammatic sketch illustrating the effects of storm wave impingement on a gently sloping shelf. Note that wave winnowing creates suspension clouds of sediment which move gravitationally down slope. Modified from Brett (1983).

northward thickening layer of coral rubble. The slope on this truncated surface is about 10° – 15°, and total relief between the top of the siltstone platform and the lowest portion of the erosion surface is about 2 m.

A series of apparent storm debris beds splay outward from a massive debris bed at the base of the slope. These layers dip gently to the northwest at a much lower angle than the slope of the platform and are separated by thickening wedges of sparsely fossiliferous shale (Fig.16). This interbedded package appears to represent a debris apron derived directly from the Staghorn platform. In this outcrop, a short but significant proximal/distal spectrum can be observed while tracing the individual debris layers.

Coral debris, apparently dislodged from the biostrome and swept off the platform edge, is represented by a thick proximal apron bordering the ramp. This unit is composed of densely packed and chaotically oriented, fragmentary and complete rugose corals (the largest up to 30 cm long). These corals, in contrast to those on the platform proper, have been heavily abraded and/or corroded. Distally, this massive rubble layer splits, within a lateral distance of 5 meters, into four discrete layers of which the lowest is thickest, most coral-rich, and contains phosphatic pebbles (Fig. 8).

Each of the three layers undergoes identical and simultaneous northward lateral changes; adjacent to the coral rubble apron, the layers are composed entirely of coral debris and are separated by thin (30 – 50 cm) wedges of gray shale containing scattered coral fragments. Traced northward, into a more distal sedimentary facies, each of the coral debris beds becomes a graded biosparite unit with a basal layer of small corals and coral fragments (Fig. 8) and an upper unit of planar to hummocky cross laminated siltstone. Each of these siltstones grade upward into gray shales which are barren aside from rare "floating" coral fragments. Approximately 200 – 300 m from the platform edge, each of the debris layers is represented by a thin (1 – 2 cm) bed of laminated siltstone; corals are generally rare, but finer skeletal debris or phosphatic pebbles commonly occur at the bed bases. The siltstone beds are penetrated by vertical traces which probably represent escape burrows (fugichnia). The intervening shale sections gradually thicken northward away from the platform toward more distal depositional regions.

The lowest coral rubble layer persists northwest of Jenny Point for about 1 km; northwest of there the siltstone becomes thin and totally burrow-riddled, making the bed rather obscure. Nevertheless, all coral debris layers may be traced to their distal expression as zones of phosphatic pebbles, shells, and small corals in silty mud matrix. The basal coral-siltstone tempestite may be traced to Fall Brook, six kilometers to the northwest; it is interpreted as a multiple episode bed, recording both nonsedimentation and rapid burial events. Clearly, from the abundance of phosphatic pebbles and worn fossils, this unit represents a long span of nondeposition, probably associated with the initial development of the Staghorn coral biostromes along the top of the platform further to the southeast. However, it also contains evidence of a "final" episode of abrupt burial. At Ten Mile Creek, well-preserved *Stereolasma* corals are found attached at their bases to phosphatic pebbles and fossil debris. Moreover, this same horizon at Fall Brook, on the west side of the lake, has yielded complete inverted, trilobites and long, articulated crinoid columns.

Thus, nearly a full spectrum of deposition types can be observed: a) amalgamated rubble beds abutting the Staghorn platform, b) parautochthonous coquinites locally transported over 100 m, c) graded siltstones, d) simple siltstones, and finally, e) mud layers which were apparently deposited up to 10 km from the platform proper.

We interpret the coral rubble beds and their lateral equivalents as the results of four major storm episodes or groups of closely-spaced storm events. During these events, storm waves impinged on the upper surface of the Staghorn coral biostromes and the siltstone margin; this wave action eroded corals, winnowed silt and mud, and swept this debris down slope to deeper, lower energy environments. Although most of the coarse coral debris accumulated at the base of the slope, smaller corals and terrigenous sediment were swept past the debris apron, forming gently sloped debris layers. Still finer-grained sediment was deposited rapidly in deeper water from suspension clouds further downslope. Hence, the Staghorn layers display much of the proposed proximality spectrum on the scale of a single outcrop. As such, these layers help corroborate the general model presented herein.

DISCUSSION

Geological processes, both cyclic and noncyclic, operate at varied temporal and geographic scales, ranging from geologically instantaneous and localized storm events to long-term global processes encompassing hundreds of thousands or millions of years. An understanding of the temporal scale of a particular geological phenomenon is critical to the reconstruction of any segment of earth history (Schindel, 1980, 1982). A major thrust of any basin analysis, including our present synthesis of the Hamilton Group in the northern Appalachian Basin, must involve the recognition and documentation of episodic sedimentary units.

Rigorous testing of depositional models, such as those presented in this volume for the Hamilton Group, ultimately requires the development of criteria for evaluating two factors: 1) relative bathymetry and 2) temporal relationships. Data pertaining to these factors are difficult, but not impossible, to obtain from ancient mudstone sequences. Commonly used bathymetric indicators (e.g., peritidal features, such as algal fossils) are rare in the Hamilton Group. Nonetheless, sedimentologic features associated with storm deposition provide a potential, but previously untapped, source of data on relative depth.

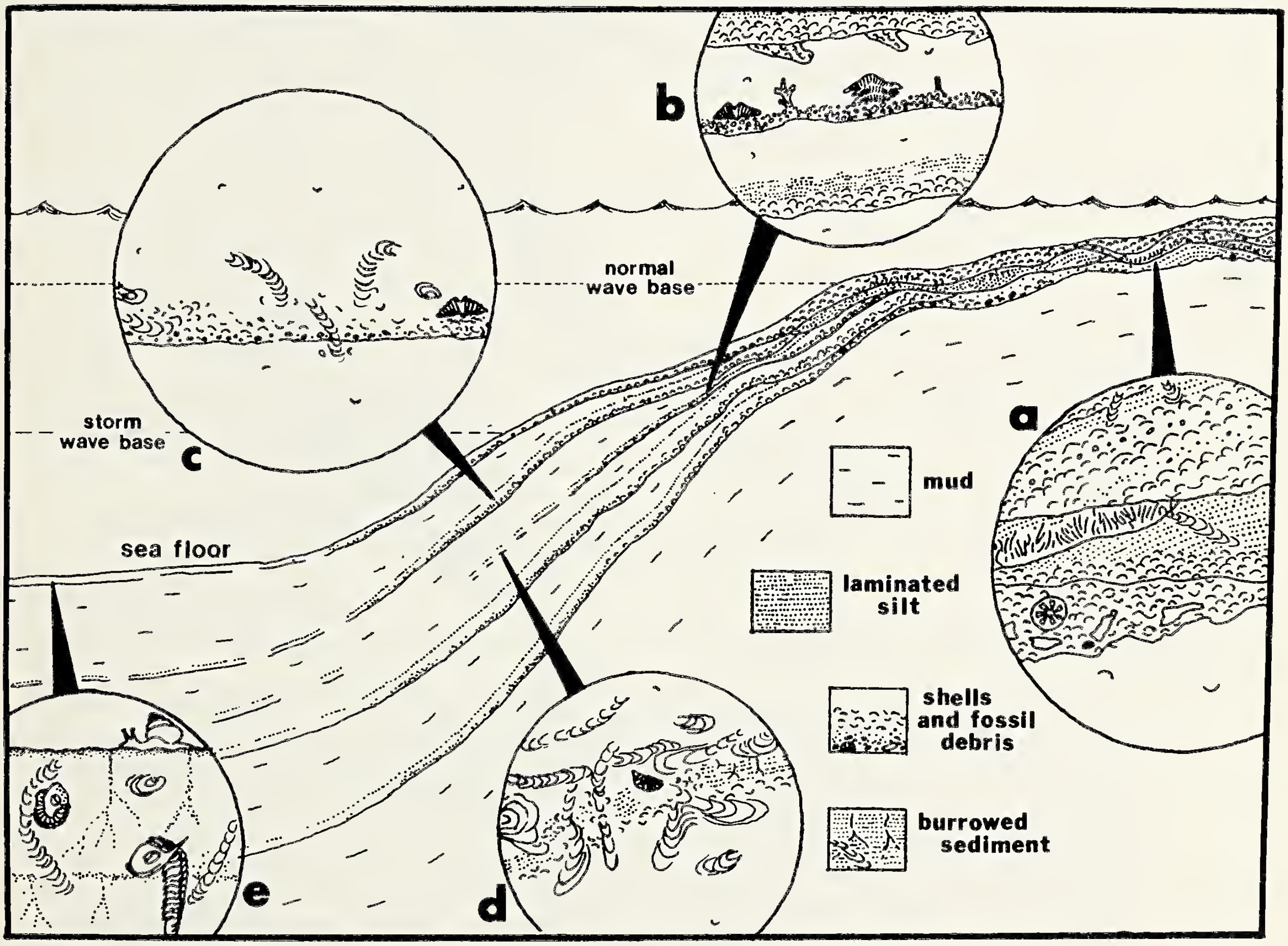


Figure 15 Characteristic depositional fabrics produced by storm action along a proximal-distal gradient, insets illustrate details as follows: a) proximal shoal facies; amalgamated tempestites; note numerous cross cutting layers displaying hummocky cross laminated fossil debris and carbonate silts; lack of mud indicates repeated winnowing; note sharply erosional base. b-d) medial tempestites displaying varying degrees of reworking; b) note sharply erosional bases of several winnowed shell lags (coquinites); lowest bed shows grading from basal shell layer into which a winnowed shell pavement has been colonized by epibionts including bryozoans and spiriferid brachiopods; these latter were subsequently buried *in situ* by a mud layer; note fillings of scoured burrows by skeletal debris in upper coquinites. c) note sharp base of winnowed shell layer and overlying, sparsely fossiliferous mudstone with burrows. d) note partial slurring of thin buried shell bed by extensive *spreiten* type bioturbation. e) distal mud tempestite; note *in situ* burial of trilobites with minor post burial bioturbation.

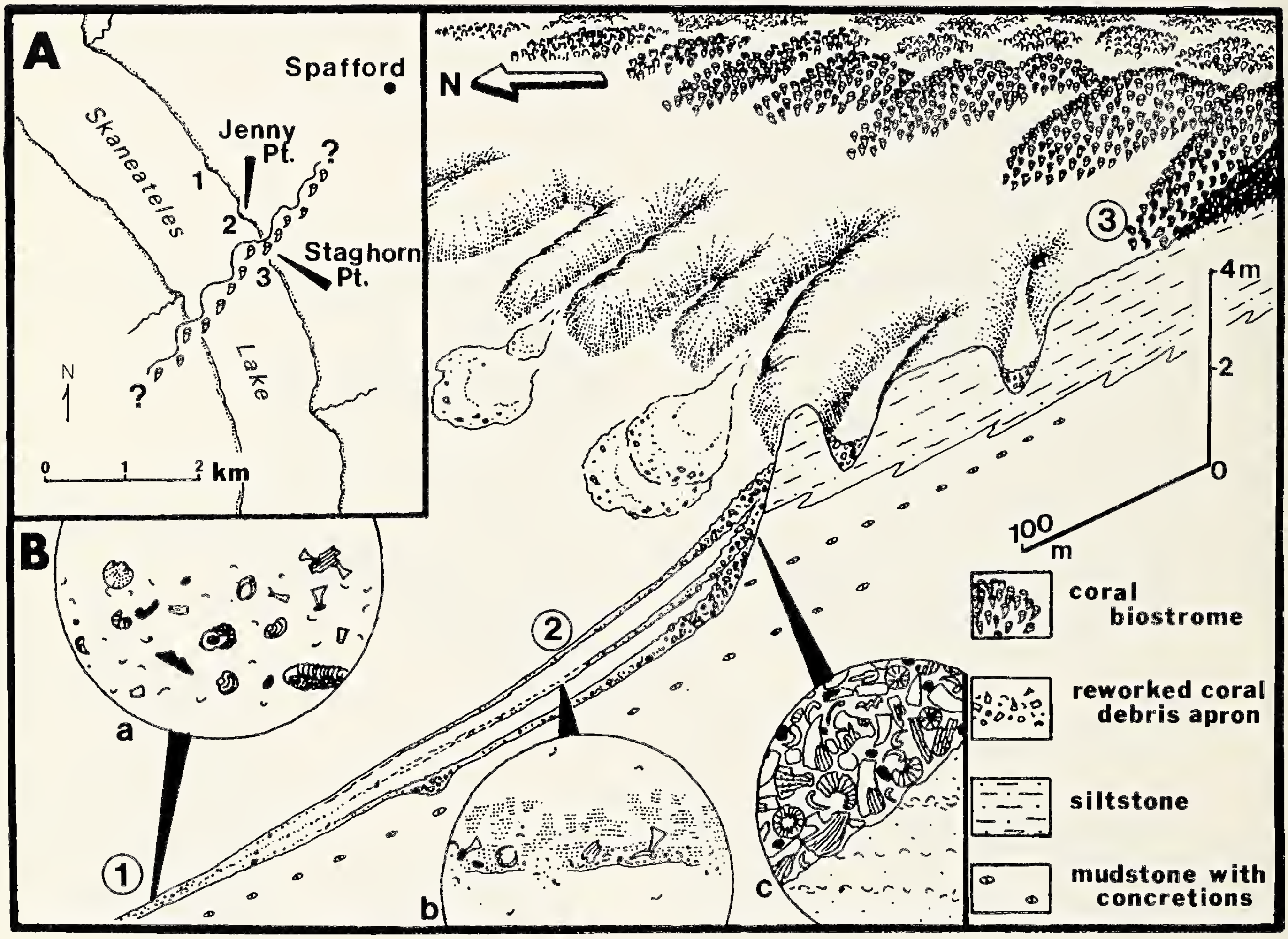


Figure 16 Proximality spectrum as illustrated by features associated with the Staghorn coral biostrome, Otisco Shale Member (Ludlowville Formation), exposed at Skaneateles Lake, Onondaga County, New York.
A) Map of a portion of Skaneateles Lake; numbers along northeast shoreline of lake correspond to positions along the reconstructed transect, based on key outcrops along Skaneateles Lake from 1 km northwest of Jenny Point to just southeast of Staghorn Point; wavy line shown diagonally across lake marks the approximate northeastern boundary of the Staghorn coral biostrome proper; the southeastern boundary of this biostrome is not exposed and is presently unknown. B) Reconstructed paleoenvironment of northwestern edge of the Staghorn coral biostrome. Coral biostrome, as exposed near location 3 (Staghorn Point), is composed of about 0.5 to 2.0 m of closely packed rugose corals (*Heliophyllum*, *Cystiphylloides*, *Siphonophrentis*) with minor claystone matrix; coral biostrome overlies a siltstone platform which caps an upward coarsening sequence about 3 – 4 m thick. Between locations 2 and 3, platform is abruptly truncated and front appears to have been a sinuous submarine escarpment; relatively few corals occur at this edge, presumably because they were eroded by episodic storm wave scour and accumulated in coarse debris aprons at the base of the slope; note three discrete debris layers which feather out and pinch together between 1 and 2.
Insets show details of sedimentary textures within storm layers: a) distal-most bed containing reworked phosphatic steinkerns and shell fragments, as well as complete brachiopods and trilobites; indicates complex burial history, including long-term nondeposition, reworking and encrustation of fossils, and final rapid burial pulse(s). b) medial tempestite with basal lag of shell and coral debris overlain by laminated siltstone. c) coarse proximal storm debris apron showing sharp erosional base overlain by amalgamated layers of coral rubble which is variably broken and corroded.

Bathymetric studies, applying the tempestite proximality model provide useful, if not essential, insight as to the nature of ancient slope and shelf environments and their associated faunas. Through the work of Aigner (1979, 1982) on the Muschelkalk and preliminary investigations of many Hamilton strata (this paper), relative depth spectra for sedimentary facies have been studied and useful generalizations drawn. However, depth studies and community analyses become most meaningful when time control is established for multiple levels in a basin margin sequence. The temporal significance of a storm-related key bed, be it complex (amalgamated) or simple (instantaneous), must first be determined or estimated. Hence, one must initially think in terms of a relative hierarchy of bed types in a given sequence before the proximality (slope-related) aspects of each bed can be fully understood. Secondly, detailed correlation of all relevant stratal information from studied sections must be undertaken so that relative changes between beds of different age and storm magnitude may be used in recognizing time-equivalent strata in adjacent sections.

For two centuries historical geologists have longed to observe ancient environments in a time-space perspective; lack of detailed correlation and knowledge of biofacies perspectives has always hampered the study of sedimentary sequences as time-lapse pictures of changing events. Recent studies, including the work of Aigner (1979, 1982, 1985) and that of Cisne and Rabe (1978), and Rabe and Cisne (1980) have come as close as any to this ideal. Cisne and Rabe, 1978) completed a comprehensive time-facies analysis (coenocorrelation) of Ordovician platform and slope communities using ash beds (i.e., bentonites). Aigner showed that proximality of tempestites can be used for similarly detailed bathymetric zonation across sedimentary basins. Examination of the proximality model (herein) and its application to the Hamilton sequence confirms its usefulness in rock sequences of varying age and character. Moreover, we submit, that where volcanic isochrons are rare or lacking, storm layers may still provide precise correlations of sedimentary sequences. We argue that sedimentary and paleoenvironmental analyses utilizing "event" horizons and proximality models as tools in correlation and interpretation will provide new insights into ancient marine environments and enhance our understanding of marine depositional systems.

ACKNOWLEDGMENTS

We thank Curt Teichert, Susan Kidwell and David Schindel for critical reviews of this paper. Margrit Gardner patiently typed various drafts of the manuscript. Acknowledgement is also made to the donors of the Petroleum Research Fund of the American Chemical Society, to National Science Foundation Grant EAR-8313103, and to the Rochester Museum and Science Center for support for field research.

REFERENCES CITED

Ager, D.V. 1973. The Nature of the Stratigraphic Record. MacMillan, London. 114 p.

__________ 1974. Storm deposits in the Jurassic of the Moroccan High Atlas. Palaeogeog. Palaeoclim. Palaeoecol. 15:83-94.

Aigner, T. 1979. Schill-Tempestite im oberen Muschelkalk (Trias, SW-Deutschland). Neues Jb. Geol. Paläeont. Abh. 157:193-217.

__________ 1982. Calcareous tempestites: storm-dominated stratification in Upper Muschelkalk limestones (Middle Trias, SW-Germany). *In* Einsele, G. and Seilacher, A., eds., Cyclic and Event Stratification. Springer Verlag, New York, Heidelberg, Berlin, p. 180-198.

__________ 1985. Storm depositional systems. Dynamic stratigraphy in modern and ancient shallow-marine sequences. Lecture notes in Earth Science 3, Springer-Verlay, New York, 174 p.

__________ and Futterer, F. 1978. Kalk-Töpfe und -Rinnen (pot and gutter casts) im Muschelkalk-Anzeiger für Wattenmeer? Neues Jb. Geol. Paläeont. Abh. 156:285-304.

__________, Hagdorn, H., and Mundlos, R. 1978. Biohermal, biostromal and storm-generated coquinas in the upper Muschelkalk. Neues Jb. Geol. Paläeont. Abh. 157:42-52.

__________ and Reineck, H.E. 1982. Proximality trends in modern storm sands from the Helgoland Bight (North Sea) and their implications for basin analysis. Senckenbergiana Marit. 14(5/6):183-215.

Ausich, W.I., Kammer, T.W., and Lane, N.G. 1979. Fossil communities of the Borden (Mississippian) delta in Indiana and northern Kentucky. J. Paleont. 53(5):1182-1196.

Ball, M.M., Shinn, E.A., and Stockman, K.W. 1967. The geologic effect of Hurricane Donna in south Florida. J. Geol. 75:583-597.

Baird, G.C. 1979. Sedimentary relationships of Portland Point and associated Middle Devonian rocks in central and western New York. New York State Mus. Bull. 433, 24 p.

__________ 1981. Submarine erosion on a gentle paleoslope: a study of two discontinuities in the New York Devonian. Lethaia 14:105-122.

__________ and Brett, C.E. 1981. Submarine discontinuities and sedimentary condensation in the upper Hamilton Group (Middle Devonian): examination of marine shelf and paleoslope deposits in the Cayuga Valley. *In* New York State Geol. Assn. Guidebook, 53rd Ann. Mtg., SUNY at Binghamton, p. 115145.

__________ and __________ 1983. Regional variation and paleontology of two coral beds in the Middle Devonian Hamilton Group of western New York. J. Paleont. 57:417-446.

__________, __________, Gray, L., and Kloc, G. in preparation. Stratigraphy, lithofacies and paleontology of the Middle Devonian Ludlowville Formation in central and western New York State.

Berner, R.A. 1968. Calcium carbonate concretions formed by the decomposition of organic matter. Science 159:195-197.

_________ 1971. Principles of Chemical Sedimentology. McGraw-Hill, Inc., New York, 240 p.

Boardman, R.S. 1960. Trepostomatous bryozoa of the Hamilton Group of New York State. U.S. Geol. Surv. Prof. Pap. 340, 87 p.

Bouma, A.H. 1962. Sedimentology of some Flysch Deposits. Elsevier, Amsterdam, 168 p.

Bowen, Z.P., Rhoads, D.C., and McAlester, A.L. 1974. Marine benthic communities of the Upper Devonian of New York. Lethaia 7:93-120.

Brenner, R.L. and Davies, D.K. 1973. Storm-generated coquinoid sandstone: genesis of high-energy marine sediments from the Upper Jurassic of Wyoming and Montana. Geol. Soc. Amer. Bull. 84:1685-1698.

Brett, C.E. 1983. Sedimentology, facies and depositional environments of the Rochester Shale (Silurian; Wenlockian) in western New York and Ontario. J. Sediment. Petrol. 53:947-971.

_________ and Baird, G.C. 1984. Comparative taphonomy: a key to interpretation of sedimentary sequences using fossil preservation. Geol. Soc. Amer. Abstracts with Programs 16(3):126.

_________ and _________ 1982. The Genesee-Moscow contact in western New York: evidence for regional erosive beveling during the late Middle Devonian. *In* New York State Geol. Assn. Guidebook, 54th Ann. Mtg., SUNY at Buffalo, p. 19-63.

_________, _________, Gray, L.M., and Savarese, M.L. 1983. Middle Devonian (Givetian) coral associations of western and central New York State. *In* Sorauf, J.E. and Oliver, W.A., Jr., eds., Silurian and Devonian Corals and Stromatoporoids of New York. Fourth Intern. Symp. on Fossil Cnidaria, Washington, D. C., p. 65-107.

_________, Dick, V.B., and Baird, G.C. (in press). Taphonomy and paleoecology of Middle Devonian (Hamilton Group) dark gray and black shales from western New York. *In* Kauffman, E.G., ed., Paleontology of Organic-Rich Strata.

_________ and Eckert, J.D. 1982. Palaeoecology of a well preserved crinoid colony from the Silurian Rochester Shale in Ontario. Life Sci. Contrib. 131, 20 p.

Brongersma-Sanders, M. 1957. Mass mortality in the sea. *In* Hedgepeth, J. W., ed., Treatise on Marine Ecology and Paleoecology, vol. 1, Ecology. Geol. Soc. Amer. Mem. 67:941-1010.

Burt, F.A. 1932. Formative processes in concretions formed about fossils as nuclei. J. Sediment. Petrol. 2:38-45.

Cant, D.J. 1980. Storm-dominated shallow marine sediments of the Arisaig Group (Silurian-Devonian) of Nova Scotia. Can. J. Earth Sci. 17:120-131.

Cisne, J.L. and Rabe, B.D. 1978. Coenocorrelation: gradient analysis of fossil communities and its applications in stratigraphy. Lethaia 11:341-364.

Clement, C.R. 1981. The occurrence and paleoecology of echinoderm assemblages in the Ludlowville Shales (Middle Devonian) of western Erie County, New York. Master's thesis, University of Cincinnati, 97 p.

Cooper, G.A. 1957. Paleoecology of the Middle Devonian of eastern and central United States. *In* Ladd, H.S., ed., Treatise on Marine Ecology and Paleoecology, Vol. 2, Paleoecology. Geol. Soc. Amer. Mem. 67, 1296 p.

Dennison, J.M. and Head, J.W. 1975. Sea level variations interpreted from the Appalachian Basin Silurian and Devonian. Amer. J. Sci. 257(1):1089-1120.

Dick, V.B. 1982. Taphonomy and depositional environments of Middle Devonian (Hamilton Group) pyritic fossil beds in western New York. Master's thesis, University of Rochester, 118 p.

_________ and Brett, C.E. 1982. Taphonomy and depositional environment of Middle Devonian (Hamilton Group) pyritic fossil horizons in western New York. Geol. Soc. Amer. Abstr. with Programs 14(1, 2):15.

Dott, R.H., Jr. 1983. Episodic sedimentation-how normal is average?how rare is rare?does it matter? J. Sediment. Petrol. 53:5-23.

_________ and Bourgeois, J. 1982. Hummocky stratification: significance of its variable bedding sequences. Geol. Soc. Amer. Bull. 93(8):663-680.

Drake, D.E. 1976. Suspended sediment transport and mud deposition on continental shelves. *In* Stanley, D.J. and Swift, D.J.P., eds., Marine Sediment Transport and Environmental Management. John Wiley, New York, 602 p.

Duke, W.L. 1982. Hummocky cross-stratification, tropical hurricanes and intense winter storms. Geol. Soc. Amer. Abstr. with Programs 14(7):478.

Eder, W. 1982. Diagenetic redistribution of carbonate, a process in forming limestone-marl alternations (Devonian and Carboniferous, Rheinisches Schiefergebirge, W. Germany). *In* Einsele, G. and Seilacher, A., eds., Cyclic and Event Stratification. Springer-Verlag, New York, Heidelberg, Berlin, p. 98-112.

Einsele, G. 1982. General remarks about the nature, occurrence, and recognition of cyclic sequences (periodites). *In* Einsele, G. and Seilacher, A. eds., Cyclic and Event Stratification. Springer-Verlag, New York, Heidelberg, Berlin, p. 3-7.

Engle, J.B. 1948. Investigation of the oyster reefs of Mississippi, Louisiana and Alabama following the hurricane of September 1947. U. S. Fish and Wildlife Service, Spec. Sci. Report 59, 70 p.

Fisher, D.W. 1951. Marcasite fauna in the Ludlowville Formation of western New York. J. Paleont. 15:61-72.

Flood, R.D. 1983. Classification of sedimentary furrows and a model for furrow initiation and evolution. Geol. Soc. Amer. Bull. 94(1):630-639.

Fürsich, F.T. 1982. Rhythmic bedding and shell bed formation in the Upper Jurassic of East Greenland. *In* Einsele, G. and Seilacher, A., eds., Cyclic and Event Stratification. Springer-Verlag, New York, Heidelberg, Berlin, p. 208-222.

Goldring, R. and Bridges, P. 1973. Sublittoral sheet sandstones. J. Sediment. Petrol. 43:736-747.

__________ and Aigner, T.W. 1982. Scour-and-fill: the significance of event separation. *In* Einsele, G. and Seilacher, A., eds., Cyclic and Event Stratification. Springer-Verlag, New York, Heidelberg, Berlin, p. 354-363.

Goldring, W. 1923. Devonian crinoids of the State of New York. New York State Mus. Mem. 16, 456 p.

Grabau, A.W. 1898-1899. Geology and paleontology of EighteenMile Creek and the lake shore sections of Erie County, New York. Buffalo Soc. Nat. Sci. Bull. 6 (Pts. 1, 2), 403 p.

Hallam, A. 1961. Brachiopod life assemblages from the marlstone rockbed of Leicestershire. Palaeontology 4:653-659.

Harno, J.C., Southard, J.B., Spearling, D.R. and Walker, R.G. 1975. Depositional environments as interpreted from primary sedimentary structures and stratification sequences. Soc. Econ. Paleont. and Mineral. Short Course No. 12, 161 p.

Heckel, P.H. 1973. Nature, origin, and significance of the Tully Limestone. Geol. Soc. Amer. Spec. Pap. 138, 244 p.

Irving, E. 1980. Phanerozoic continental drift. Physics of the Earth and Planetary Interiors 24:197-204.

Kesling, R.V. and Chilman, R.B. 1975. Strata and Megafossils of the Middle Devonian Silica Formation. Univ. Mich. Mus. of Paleont., Papers on Paleontology 8, 408 p.

Klapper, G. 1981. Review of New York conodont biostratigraphy. *In* Oliver, W.A. and Klapper, G., eds., Devonian Biostratigraphy of New York. Intl. Union of Geol. Sci., Subcomm. on Devonian Strat., Washington, D.C., p. 57-66.

Kloc, G. 1983. Stratigraphic distribution of ammonoids from the Middle Devonian Ludlowville Formation in New York. Master's thesis, SUNY at Buffalo, 73 p.

Kidwell, S.M. and Jablonski, D. 1983. Taphonomic feedback: ecological consequences of shell accumulation. *In* Tevesz, M.J.S. and McCall, P.L., eds., Biotic Interactions in Recent and Fossil Benthic Communities. Plenum Press, New York, p. 195-248.

Kranz, D.M. 1974a. Anastrophic burial of bivalves and its paleoecological significance. J. Geol. 82:237-265.

__________ 1974b. Computer simulation of fossil assemblage formation under conditions of anastrophic burial. J. Paleont. 48:800-808.

Kreisa, R.D. 1981. Storm-generated sedimentary structures in subtidal marine facies with examples from the Middle and Upper Ordovician of southwestern Virginia. J. Sediment. Petrol. 51:823-848.

__________ and Bambach, R.K. 1982. The role of storm processes in generating shell beds in Paleozoic shelf environments. *In* Einsele, G. and Seilacher, A., eds., Cyclic and Event Stratification. Springer-Verlag, New York, Heidelberg, Berlin, p. 200-207.

Kumar, N. and Sanders, J.E. 1976. Characteristics of shoreface storm deposits: modern and ancient examples. J. Sediment. Petrol. 46:145-162.

Landing, E. and Brett, C.E., in press. Trace fossils and regional significance of a Middle Devonian (Givetian) disconformity in southwestern Ontario. J. Paleo. 60.

Larson, D.W. and Rhodes, D.C. 1983. The evolution of infaunal communities and sedimentary fabrics. *In* Tevesz, M.J.S. and McCall, P.L., eds., Biotic Interactions in Recent and Fossil Benthic Communities. Plenum Press, New York, p. 627-648.

Liddell, W.D. 1975. Recent crinoid biostratinomy. Geol. Soc. Amer. Abstr. with Programs 7(7):1169.

Linsley, R.M. 1972. Stop 2: 'Pierceville Quarry' (Bradley Brook Quarry); paleoecology of the Ludlowville Formation, Hamilton Group. *In* McLelland, J., ed., New York State Geol. Assn. Guidebook, 44th Ann. Mtg., Colgate Univ. and Utica College, p. F8-F18.

McKnight, D.G. 1969. A recent, possible catastrophic burial in a marine molluscan community. New Zealand J. Mar. and Fresh. Res. 3:177-179.

Marsaglia, K.M. and DeVries-Klein, G. 1983. The paleogeography of Paleozoic and Mesozoic storm depositional systems. J. Geol. 91:117-142.

Meyer, D.L. 1971. Post-mortem disarticulation of Recent crinoids and ophiuroids under natural conditions. Geol. Soc. Amer. Abstr. with Programs 3(7):645.

Miller, K.B. 1982. Paleomicroecologic analysis of depositional environments and sequences, "*Pleurodictyum* Zone," Ludlowville Formation, Middle Devonian of western New York. Master's thesis, SUNY at Binghamton, 131 p.

Miller, M.F. and Rehmer, J.E. 1982. Using biogenic structures to interpret sharp lithologic boundaries: an example from the Lower Devonian of New York. J. Sediment. Petrol. 52(3):887-897.

Oliver, W.A., Jr. 1951. Middle Devonian coral beds of central New York. Amer. J. Sci. 249(10):705-728.

__________ 1976. Biogeography of Devonian rugose corals. J. Paleont. 50:365-373.

Plotnick, R.E. 1984. Biostratinomy and early diagenesis of modern arthropods. Geol. Soc. Amer. Abstr. with Programs 16(3):186.

Potter, P.E., Maynard, J.B., and Pryor, W.A. 1980. Sedimentology of Shale. Springer-Verlag, New York, Heidelberg, Berlin, 306 p.

Rabe, B.D. and Cisne, J.L. 1980. Chronostratigraphic accuracy of Ordovician stratigraphic correlation. Lethaia 13:109-118.

Raiswell, R. 1976. The microbiological formation of carbonate concretions in the upper Lias of NE England. Chem. Geol. 18:227-244.

Reineck, H.E. and Singh, I.B. 1972. Genesis of laminated sand and graded rhythmites in storm sand layers of shelf mud. Sedimentology 18:123-128.

Rhoads, D.C. and Morse, J.W. 1971. Evolutionary and ecologic significance of oxygen-deficient marine basins. Lethaia 4:413-428.

Rich, J.L. 1951. Three critical environments of deposition, and criteria for recognition of rocks deposited in each of them. Geol. Soc. Amer. Bull. 62:1-20.

Rickard, L.V. 1981. The Devonian System of New York State. *In* Oliver, W.A., Jr. and Klapper, G., eds., Devonian Biostratigraphy of New York. Intl. Union Geol. Sci., Subcom. on Devonian Strat. Part 1, Washington, D.C., p. 5-22.

Rodolfo, K.S., Buss, B.A. and Pilkey, O.H. 1971. Suspended sediment increase due to Hurricane Gerde in continental shelf waters off Cape Lookout, North Carolina. J. Sediment. Petrol. 41:1121-1125.

Sass, E. and Kolodny, Y. 1972. Stable isotopes, chemistry and petrology of carbonate concretions (Mishash Formation, Israel). Chem. Geol. 10:261-286.

Savarese, M., Gray, L.M. and Brett, C.E. this volume. Faunal and lithological cyclicity in the Centerfield Member (Middle Devonian: Hamilton Group) of western New York: A reinterpretation of depositional history).

Schindel, D.E. 1980. Microstratigraphic sampling and the limits of paleontological resolution. Paleobiology 6:408-426.

__________ 1982. Resolution analysis: a new approach to the gaps in the fossil record. Paleobiology 8:340-353.

Seilacher, A. 1970. Begriff und Bedeutung der Fossil-Lagerstatten. Neues Jb. Geol. Palaeontol. Monatsh. 1970:34-39.

__________ 1982. General remarks about event deposits. *In* Einsele, G. and Seilacher, A., eds., Cyclic and Event Stratification. Springer-Verlag, New York, Heidelberg, Berlin, p. 161-174.

Sepkoski, J.J., Jr. 1982. Flat pebble conglomerates, storm deposits, and the Cambrian bottom fauna. *In* Einsele, G. and Seilacher, A., eds., Cyclic and Event Stratification. Springer-Verlag, New York, Heidelberg, Berlin, p. 371-385.

Smith, B. 1912. Observations on the structure of some coral beds in the Hamilton shale. Acad. Nat. Sci. Phila. Proc. 1912:446-454.

__________ 1935. Geology and mineral resources of the Skaneateles Quadrangle. New York State Museum Bull. 300, 120 p.

Speyer, S.E. 1983. Trilobite clustering in the Hamilton Group of New York State. Master's thesis, The University of Rochester, 113 p.

__________ and Brett, C.E. 1982. Intraspecific clustering in Middle Devonian trilobites; the "coterie phenomenon." Geol. Soc. Amer. Abstr. with Programs 14(7):623.

__________ and __________ 1984. Comparative taphonomy of Middle Devonian trilobite beds. Geol. Soc. Amer. Abstr. with Programs 16(3):198.

__________ and __________ 1985. Clustered trilobite assemblages in the Middle Devonian Hamilton Group. Lethaia 18:85-103.

Stitt, J.H. 1976. Functional morphology and life habits of the Late Cambrian trilobite *Stenopilus pronus* Raymond. J. Paleont. 50:561-576.

Tarr, W.A. 1921. Syngenetic origin of concretions in shale. Geol. Soc. Amer. Bull. 32:373-384.

Thayer, C.W. 1981. Sediment-mediated biological disturbance and the evolution of marine benthos. *In* Tevesz, M.J.S. and McCall, P.L., eds., Biotic Interactions in Recent and Fossil Benthic Communities. Plenum Press, New York, p. 480-625.

Waage, K.M. 1964. Origin of repeated fossiliferous concretion layers in the Fox Hills Fm. (Cret., S. D.). Kansas Geol. Surv. Bull. 169:541-563.

Walker, R. 1967. Turbidite sedimentary structures and their relationship to proximal and distal depositional environments. J. Sediment. Petrol. 37:25-43.

Walther, M. 1982. A contribution to the origin of limestone shale sequences. *In* Einsele, G. and Seilacher, A., eds., Cyclic and Event Sedimentation. Springer-Verlag, New York, Heidelberg, Berlin, p. 113-120.

Weeks, L.G. 1953. Environment and mode of origin and facies relationships of carbonate concretions in shales. J. Sediment. Petrol. 23:162-173.

__________ 1957. Origin of carbonate concretions in shales, Magdalena Valley, Columbia. Geol. Soc. Amer. Bull. 68:95102.

Zangerl, R., Woodland, B.B., Richardson, E.S., and Zachary, D.L. 1969. Early diagenetic phenomena in the Fayetteville Black Shale (Mississippian) at Arkansas. Sediment. Geol. 3:87120.

SMITHSONIAN INSTITUTION LIBRARIES
3 9088 01300 8966

ISSN 0278-3355
ISBN 1-55557-003-8